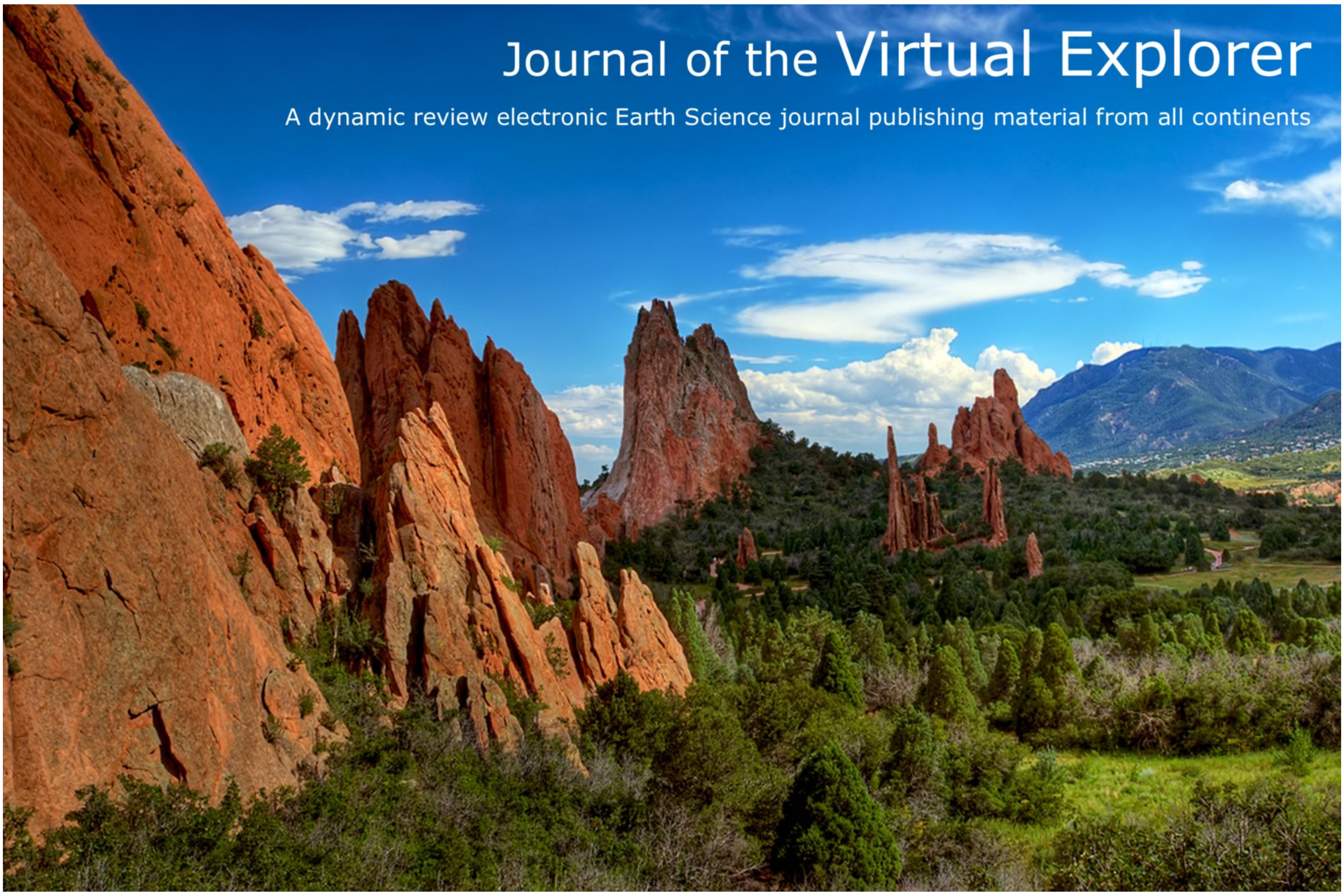

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Tectonics, magmatism and geodynamics of Italy: What we know and what we imagine

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Abstract: The Cenozoic geological evolution of the Italian area is characterized by the formation of two major mountain chains - the Alps to the north and the Apennines throughout the peninsula – plus the opening of two oceanic basins (the Ligurian-Provençal and the Tyrrhenian Sea). Associated with the formation of these two belts, a volumetrically important and chemically complex magmatic activity developed. The Alps and the Apennines show very different styles of evolution: the Alps display double-verging growth, with the involvement of large volumes of basement and the exhumation of metamorphic rocks (thick-skinned tectonics). On the other hand, the Apennines are a single-verging belt, mostly characterized by thin-skinned tectonics and associated to a radial "eastward" translation (coupled to extensional tectonics in the Ligurian-Provençal, Tyrrhenian, and western Apennines areas). The Apennines generated an arc from the northern Apennines down to Sicily, possibly merging with the Maghrebides along the northern Africa coast. The different evolution of the Alpine and Apenninic belts is mirrored by the different geometry of the respective foredeep or foreland basins (shallow in the Alps and deep in the Apennines), as recorded also by the dip of the foreland monocline (shallow in the Alps, 2-4°, and steeper in the Apennines, 6-15°).

The paradox is evident: the higher the belt, the thinner the foreland basin. The Alps consist of rocks belonging to the continental margings of the European and Adriatic-African plates, as well as remnants of Mesozoic intervening ocean(s). On the other hand, the Apennines, with the exception of the Calabro-Peloritani arc and other scattered basement outcrops, are mainly made up of rocks of Adriatic origin (Mesozoic Laziale-Abruzzese and Apulian carbonate platforms plus basinal successions), with subordinate ophiolites. From a magmatological point of view, the Alpine magmatism is essentially concentrated in a relatively narrow area, the so-called Insubric Lineament and in a relatively short time (mostly ~32-24 Ma). On the other hand, the Apennines-related igneous activity spans a larger time range (essentially from 22 Ma to Present), with several peaks in magma production. This magmatism took place over a much wider area, characterized by variable lithospheric thickness, Moho age and depth.

On the basis of thermo-tectonic, magmatological, and plate-kinematics constraints, a geodynamic evolutionary model of the Italian area is proposed. We suggest that three subduction zones have been active and have consumed oceanic and, partially, continental lithosphere: the Alpine subduction zone, with the European plate under-thrusting the Adriatic microplate; the Apenninic subduction zone, with the ancient (Mesozoic?) Ionian/Mesogean oceanic lithosphere and the Adriatic micro-plate under-thrusting westward the European plate; and the Dinaric subduction zone, with the Adriatic micropate under-thrusting northeastward the European plate. Such a geodynamic scenario is summarised in a movie, spanning the last 50 Ma.

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Introduction

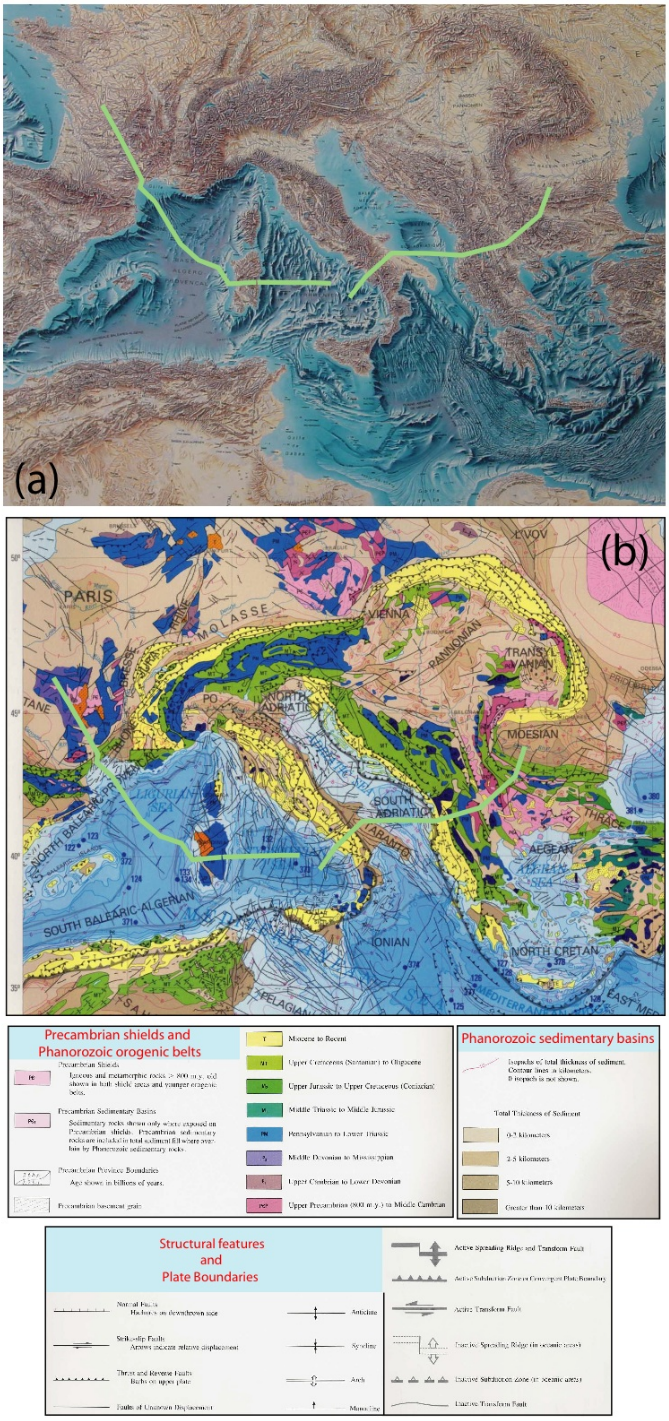
The Italian area is characterized by strong areal and temporal variability of tectonic and magmatic signatures. This variability affects also the entire Central-Western Mediterranean (Fig. 1 and 2) and challenges the applicability of classical plate tectonics models (e.g., Lustrino *et al.*, 2011). This variability, coupled to the incompleteness of geological and geophysical observables and to their non-unique interpretation, resulted in the proposal of several, often contrasting, geodynamic scenarios of evolution. In this work we use thermo-tectonic, magmatological, and plate-kinematics constraints (what we know) to propose our best-guess geodynamic evolution (what we imagine) from the Late Cretaceous to the Present. Alternative geodynamic scenarios proposed in the literature are briefly discussed. The (Permian-Early Cretaceous) rifting and drifting stages of the Alpine cycle will not be treated in this work, although the influence of the paleogeography inherited during these stages on the architecture of the contractional and extensional features developed in the Tertiary are taken into account. A movie illustrating our idea of evolution of the western-central Mediterranean area during the last 50 Ma is provided, together with a frame-by-frame pdf file containing additional information and some notes, and can be downloaded from the Journal of Virtual Explorer repository system.

What we know

Plate kinematics

Africa started to move against the southern European paleo-continental margin during Early Cretaceous, when the first rifting stages in southern Pangea gave room to oceanization processes in the southern Atlantic Ocean (e.g., Dercourt *et al.*, 1986; Eagles, 2007; and references therein). The Tethyan Ocean, located between European and Africa continental plates, started to be consumed with different subduction polarities and complex geometries, also due to the presence of still poorly constrained and debated continental micro-plates in between (Iberia, Adria, AlKaPeCa, Briançonnais, Sesia-Lanzo, and so on). The vectors of the first stage Africa-Europe relative motions are not known in detail and different models have been proposed in the literature. During the Paleogene, a N-NE motion of Africa against Europe (assumed to be fixed) was proposed (Dewey *et al.*, 1989; Mazzoli and Helman, 1994; Rosenbaum *et al.*, 2002a).

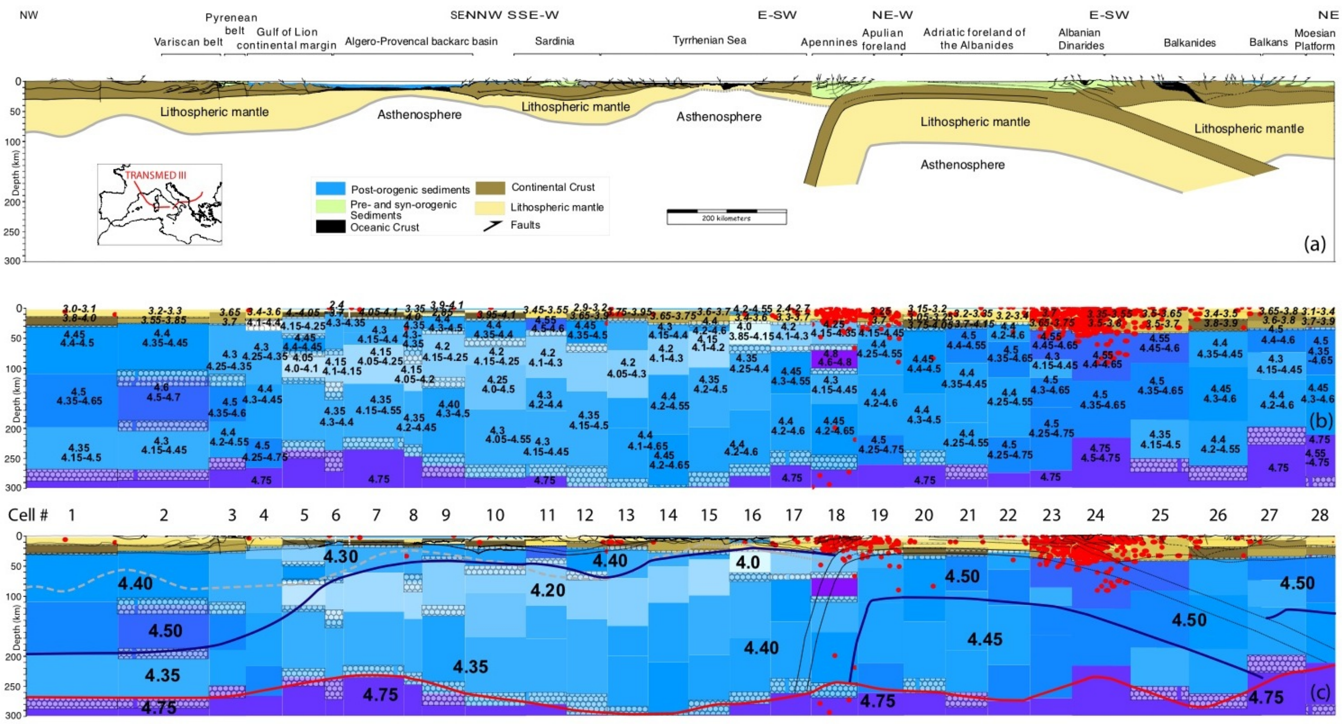
Figure 1. Topography-bathymetry and geological map of the Mediterranean area.



(a) Topography-bathymetry (after Tanguy de Rémur) and (b) geological map (After EXXON Prooduction Research Company, 1994) of the Mediterranean area. The trace of the cross section of Fig. 2 is shown.

A major change happened during the Eocene, when convergence rates reduced to <1 cm/yr and the convergence direction changed to N-NW (e.g., Rosenbaum *et al.*, 2002a, 2002b, and references therein). Recent space geodesy data confirm this main frame, where Africa has about 4-5 mm/yr of N-S component of convergence relative to Europe (e.g., Devoti *et al.*, 2008). To complicate the scenario, geodynamic evidence suggests that Africa and the intervening microplates have been affected by simultaneous rotations (i.e., rotation and sub-rotations, Cuffaro *et al.*, 2008) since at least the Early Oligocene. This represents a key kinematic constraint to understand the evolution of tectonic settings in the Mediterranean Area.

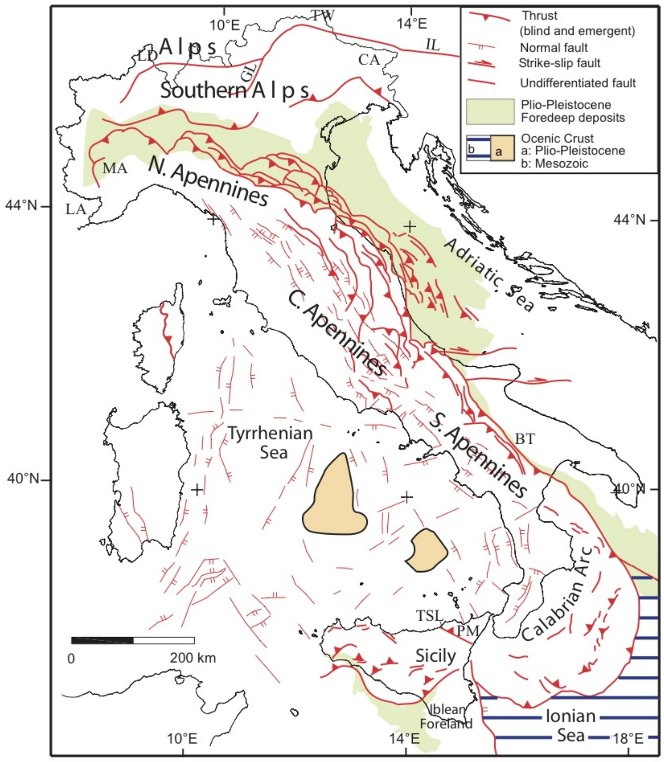
Figure 2. Lithospheric scale cross-section through the Central Mediterranean



a) Lithospheric scale cross-section through the Central Mediterranean (simplified and redrawn from the TRANSMED III geotraverse; Carminati *et al.*, 2004). (b) Lithosphere-asthenosphere system along the trace of the TRANSMED III geotraverse: the tomographic cross section was obtained from selected solutions and related seismicity (body waves magnitude greater or equal to 3.0). The VS and its range of variability in km/s are printed on each layer. Numbers in italic denote the velocities in the crustal layers. The hypocenters are denoted by dots. (c) Overlap between the TRANSMED III regional cross-sections and the Vs tomography (after Panza *et al.*, 2007).

The nature, size and kinematics of the Adriatic area is also debated. Paleomagnetism, propagation of transverse waves refracted below the Mohorovicic discontinuity, paleoclimatology and palinology suggest that Adria belongs structurally and kinematically to Africa (e.g., Mantovani *et al.*, 1990, Channell, 1996; Muttoni *et al.*, 1996; Mele, 2001), whereas historical seismicity, geodetic and seismic evidence suggest that Adria is now an independent microplate within the Africa-Eurasia plate boundary zone (Anderson and Jackson, 1987; Nocquet and Calais, 2003). In the following, Adria is considered, as a first approximation, separated from Africa during the time of opening of the Ionian oceanic basin, whose age is still poorly constrained. (Fig. 3; Catalano *et al.*, 2001). Nevertheless, for sake of simplicity and lack of better constraints, the Adriatic plate is assumed to have moved coherently with Africa in the movie attached to this article.

Figure 3. Simplified structural map of Italy (modified from Consiglio Nazionale delle Ricerche, 1992; Scrocca, 2006).



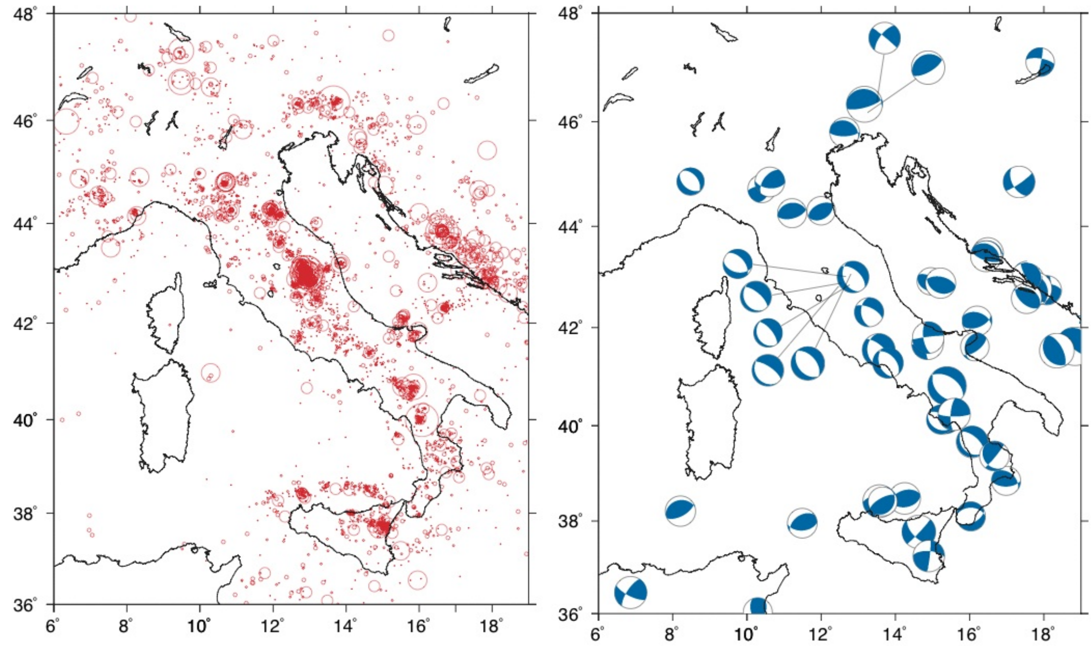
LA: Ligurian Alps; MA: Monferrato Apennines; IL: Insubric Line; CA: Carnian Alps; GL: Giudicarie Line; LD: Lepontine dome; TW: Tauern Window; PM: Peloritani Mountains; BT: Bradanic trough; TSL: Taormina-Sangineto Line.

Tectonic and magmatic constraints

Italy is a tectonically (Fig. 3) and magmatologically active region, as testified by the distribution of earthquakes (Fig. 4) and igneous activity. Earthquakes show prevalent compressional focal mechanisms at the fronts of Alps and Apennines and extensional mechanisms along the Apennines backbone. This picture is confirmed by in-situ stress measurements (e.g., borehole breakouts; Montone *et al.*, 2004). As it will be discussed later, Alps and Apennines show markedly different characters (Carminati *et al.*, 2004a). The Apennines tectonics partly overlap the Alpine structures in the Ligurian Alps-Monferrato Apennines (Fig. 3; Laubscher, 1971; Schumacher and Laubscher, 1996; Molli et al., this volume). The different evolution of the Alpine and Apenninic belts is mirrored by the different geometry of the respective foredeep basins (shallow in the Alps and deep in the Apennines), strongly controlled by the dip of the foreland monocline (shallow in the Alps and steeper in the Apennines; Mariotti and Doglioni, 2000). The geometry of the compressional structures is mainly controlled by active geodynamic processes and paleogeography (distribution of structural highs and lows, inherited from the Mesozoic rifting). Salients (where the tectonic front is more advanced) and steeply dipping regional monoclines (e.g., the top of the basement) are normally associated with Mesozoic basins, characterized by thick successions and deep basal *decollements*. Recesses (where the front is less advanced) and less steep regional monoclines are rather associated with inherited Mesozoic structural highs. Some of the geophysical constraints (heat flow distribution, crustal and lithospheric thickness, Bouguer anomaly) available for the Italian area are summarized in Figs. 5 and 6.

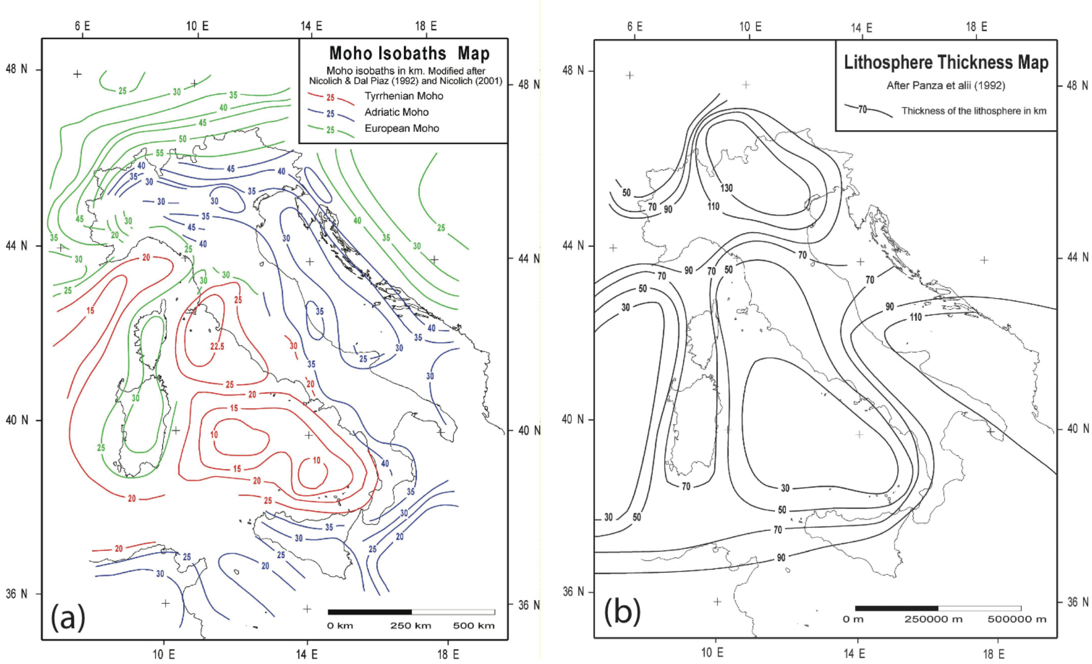
Magmatism is rather controlled by lithospheric scale discontinuities, as clearly evidenced by the roughly linear alignment of the Cenozoic igneous rocks of the Alps, and by active geodynamic processes, as observed in Sardinia, Sicily, along the Italian peninsula, the south-eastern Mediterranean and the Sicily Channel.

Figure 4. Earthquake hypocentres and Centroid moment sensor solutions



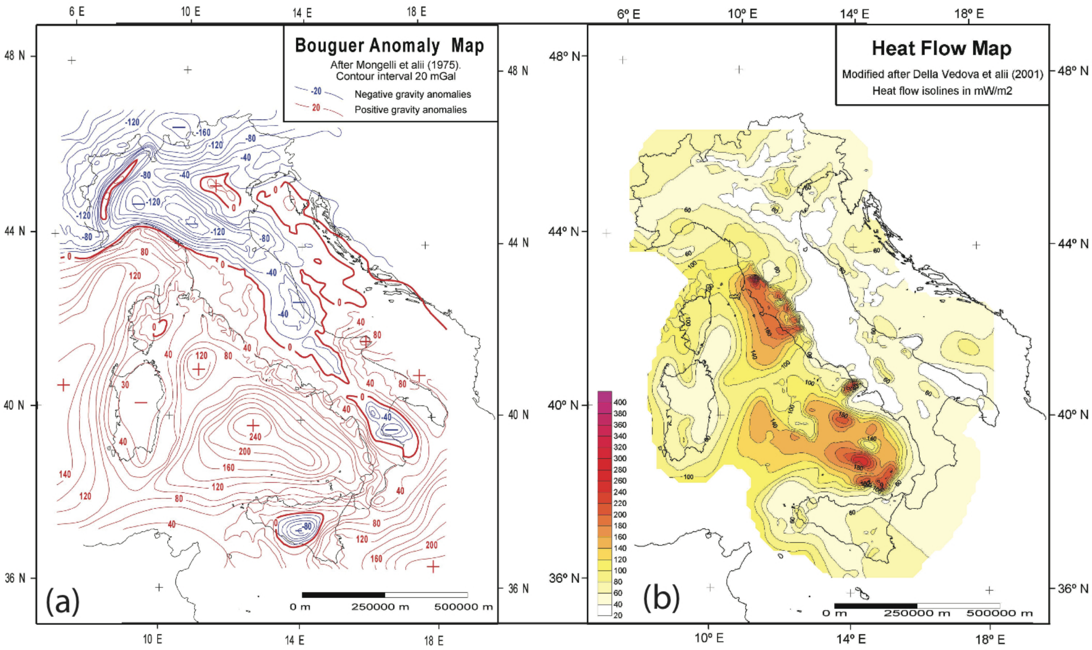
a) Earthquake hypocentres recorded by the Italian telemetered seismic network (ITSN) in the years 1989-1998 whose magnitudes are greater than 2.4. Data from INGV (http://www.ingv.it). b) Centroid moment tensor solutions available for the Italian area. Only earthquakes shallower than 40 km were selected. Data from the Harvard CMT Catalog (http://www.seismology.harvard.edu/CMTsearch.html).

Figure 5. Crustal and lithospheric thickness



(a) Crustal and (b) lithospheric thickness, (after Nicolich and Dal Piaz, 1992; Nicolich, 2001; Panza *et al.*, 1992, in Scrocca *et al.*, 2003).

Figure 6. Bouguer anomalies and heat flow density



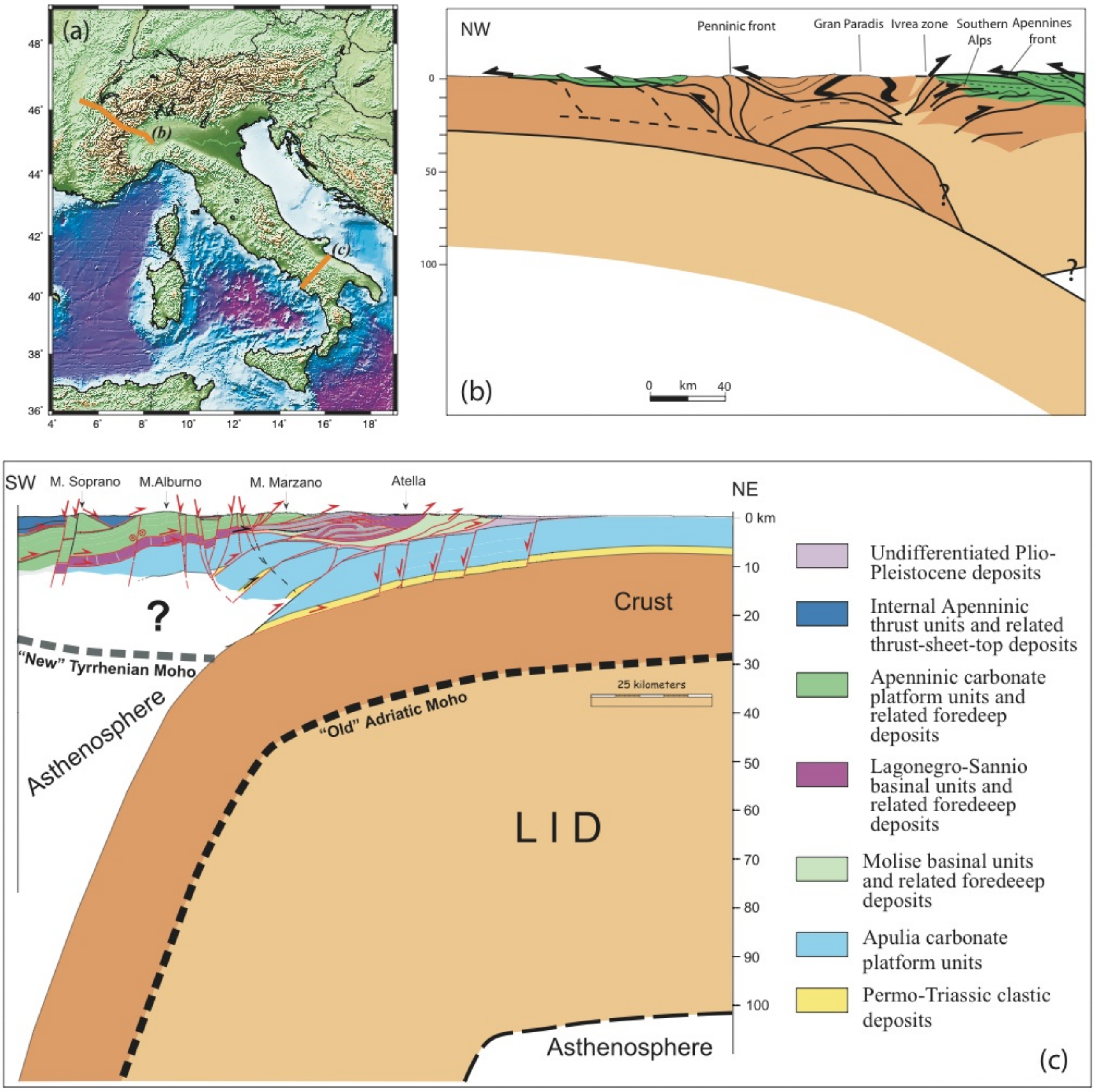
(a) Bouguer anomalies and (b) heat flow density of the Italian area (after Mongelli, 1975; Cataldi *et al.*, 1995; Della Vedova *et al.*, 2001).

Alps

The Alps (Figs. 3 and 7) show a double-vergence growth, with the involvement of large volumes of basement and the exhumation of metamorphic rocks. These rocks belong to the European (Helvetic domain in the literature), oceanic (Penninic domain), and African realms. The Africa-derived rocks belong actually to the Adriatic (also known as Apulian) micro-plate and are called Austroalpine and Southalpine domains, if cropping out south and north of the Insubric Line respectively. The western and central Alps are characterized by a discontinuous belt of crystalline massifs, which disappear eastward in the Pannonian Basin, being buried under thousands of meters of Neogene sediments (Fügenschuh and Schmid, 2003). The foredeep of the Alps has the deepest depocenter to the north of the central Alps (Fig. 8). Lithospheric-scale cross sections through the Central and Western Alps are provided in Figs. 7. The Southern Alps (Fig. 7) are a south-verging thick-skinned fold-and-thrust belt composed of basement and cover rocks belonging to the Adriatic microplate, which were affected at most by very low-grade metamorphism during the Alpine orogeny (e.g., Carminati *et al.*, 1997 and references therein). The Southern Alps are separated from the north-verging Alpine, mainly metamorphic, orogen by the Insubric Line, a dextral transpressive mylonitic-to-cataclastic belt active since Oligocene (e.g., Rosenberg, 2004 and refences therein). In the Alps, water divide and highest peaks are roughly coincident (Fig. 9).

The onset of compressional deformation in the Alps during Late Cretaceous-Paleocene drove to Europe-verging nappe formation associated with high-pressure metamorphism (eclogites and eclogite-facies continental rocks; Dal Piaz *et al.*, 1972; Spalla *et al.*, 1996; Berger and Bousquet, 2008) and to deposition of coeval flysch (Polino *et al.*, 1990). High P/T (e.g., Ernst, 1971) and ultra-high pressure (e.g., Chopin, 1984) metamorphic rocks and flysch deposits are mainly found in the Penninic accretionary prism and in the Alpine Corsica units (e.g., Spalla *et al.*, 1996; Molli *et al.*, 2008 for reviews). Compressive deformations are recorded, at this stage, also in the Southern Alps (e.g., Doglioni and Bosellini, 1987; Carminati and Siletto, 1997; Carminati *et al.*, 1997). These deformations were interpreted as the result of the oceanic subduction stage.

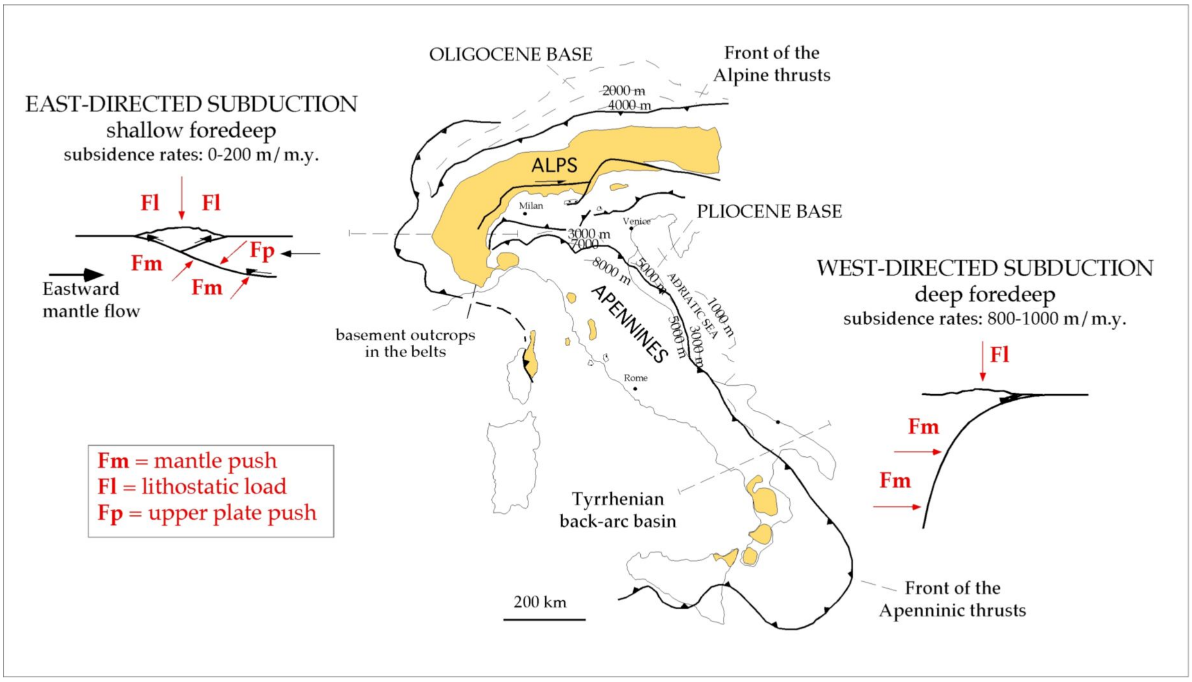
Figure 7. Lithospheric scale cross sections through Alps and Apennines



Lithospheric scale cross sections through the (b) western Alps (redrawn and modified after Roure *et al.*, 1990b) and (c) southern Apennines (modified after Scrocca *et al.*, 2005). The traces of the sections are shown in panel a. Panel c also shows the vertical motions along the transect and their sources.

Following the diachronous Eocene-Oligocene? Adria-Europe collision, north of the Insubric Line the Penninic nappes were thrust north onto the foreland forcing the detachment of sediments from their original substratum (the European plate), leading to the formation of the Helvetic nappes. Early Eocene ages for the collision in the Central-Eastern Alps are supported by the Paleocene-Early Eocene age of the youger flysch (sediments interpreted as deposited in trench environment) deposits (Stampfli et al., 1998) and by the 42-40 Ma Ar-Ar ages for the pressure peak metamorphism in the rocks belonging to the subducted oceanic realm (Wiederkehr et al., 2009). Such pressure peak is expected to postdate by a few Ma the subduction of the rocks. On the contrary, in the Western Alps, the youngest flysches were dated as Late Eocene to possibly Early Oligocene in age (Bagnoud et al., 1998). Young ages (Late Eocene-Oligocene) for collision are also supported by the young ages (as young as 35 Ma) of ductile shear zones developed at high-pressure metamorphic (i.e., tipically related to subduction preocesses) conditions, obtained from rocks belonging to the oceanic and to the European realms (Beltrando et al., 2010). Post-collisional shortening has been evaluated to be about 200 km in a N-S transect crossing central Switzerland (Schmid *et al.*, 1995). Contemporaneous southward thrusting and associated folding occurred south of the Insubric Line, and shortening amounted to ~55 km in the central Southern Alps (Schönborn, 1992) and ~40-55 km in the Eastern Southern Alps (Castellarin *et al.*, 2006). From the Miocene onwards, shortening in the Southern Alps outweighted the one occurred north of the Insubric Line. In the eastern Southern Alps and in the Carnian Alps (Fig. 3), interference occurred between the NW-SE-trending thrusts related to the pre-Neogene (Cretaceous-Paleogene?) forebelt of the Dinarides subduction and the ~E-W Neogene-Quaternary thrusts related to the structuration of the retrobelt of the Alp (i.e., the Southern Alps; Doglioni, 1987; Merlini *et al.*, 2002; Ponton *et al.*, 2010).

Figure 8. Alps and Apennines subduction zones and orogens



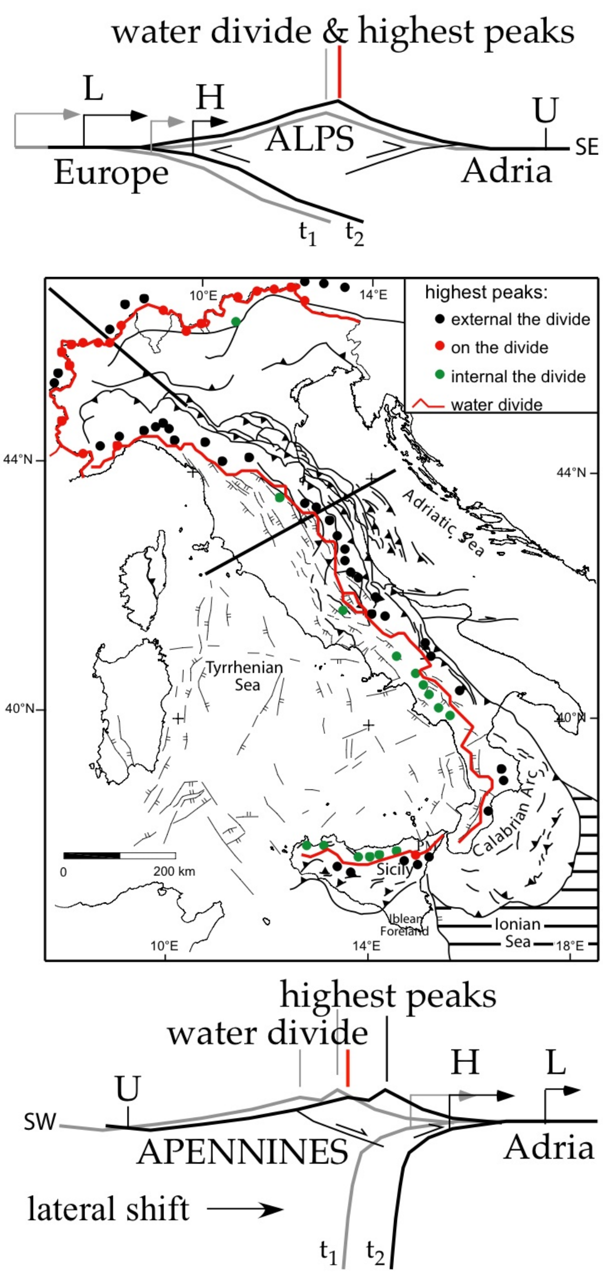
Alps and Apennines are orogens with very different characters, which may have resulted from markedly different subduction zones. The Alps have a shallow foredeep (the deepest depocenter has the Oligocene base at about 4 km), broad outcrops of metamorphic rocks (in yellow), and high structural and morphologic relief. The Apennines have deep foredeep (the deepest depocenter has the Pliocene base at about 8.5 km), few outcrops of basement rocks (partly inherited from earlier Alpine phase), low structural and morphologic elevation, and the Tyrrhenian back-arc basin. These differences mimic asymmetries of Pacific east-directed Chilean and west-directed Marianas subductions. For E-NE-directed subduction foredeep and trench, the origin may be interpreted as controlled by the lithostatic load and the downward component of upper plate push, minus the upward component of the "eastward" mantle flow (upper left panel). For the W-directed subduction where subsidence rate is much faster (>1 mm/yr), the foredeep origin may rather be interpreted as due to the horizontal mantle push and the lithostatic load [panel in the lower right (modified after Doglioni, 1994)].

In the central Southern Alps the post-Oligocene shortening was mainly accommodated by thrusting and folding along the Milano Belt, sealed by the ~6 Myr-old Messinian unconformity and buried under the Pliocene-Pleistocene sediments of the Po Plain (Figs. 3 and 7; Pieri and Groppi, 1981). Pliocene deformation was accomplished by out-of-sequence (i.e., positioned in more internal positions with respect to the outernmost front) thrusts (Schönborn, 1992). East of the Giudicarie Line (Fig. 3), that represents a sinistral transpressional tectonic lineament connected to the larger Insubric Line, the front of the Southern Alps crops out. With the exception of the NW-SE trending front of the Carnian Prealps (Merlini *et al.*, 2002), such a front is segmented by transfer faults (e.g., Doglioni, 1992). In the eastern portions of the Southern Alps, N-S shortening is well documented from the Miocene to the present with an in-sequence thrust arrangement (Castellarin and Vai, 1981; Benedetti *et al.*, 2000).

In the Alps (north of the Insubric Line), collision-related (Barrovian-type) metamorphism (Dal Piaz *et al.*, 1972; Dal Piaz, 2001) has been recognized in the external zones north of the axial part of the chain, as well as in the central Alps (Lepontine dome) and in the Tauern window of the eastern Alps (Fig. 3 and 7; Bousquet *et al.*, 2008). In the external zones, where only lower greenschist facies conditions were reached, metamorphism resulted from burial associated with nappe emplacement, thrusting and folding (e.g., Burkhard and Goy-Eggenberger 2001; Ferreiro-Maehlmann, 2001). The Lepontine area is characterized by widespread amphibolite facies metamorphism, but granulite facies conditions and migmatization were locally reached (Berger *et al.*, 2005). The Lepontine metamorphism progressively decreases from upper-amphibolite facies conditions in the south (where it is truncated by the Insubric line, juxtaposing such high-grade rocks with rocks of the southern Alps that at most experienced Alpine anchimetamorphism) to greenschist facies conditions moving to the north (Streckeisen *et al.* 1974; Todd and Engi 1997; Engi *et al.*, 2004; Berger *et al.*, 2005).

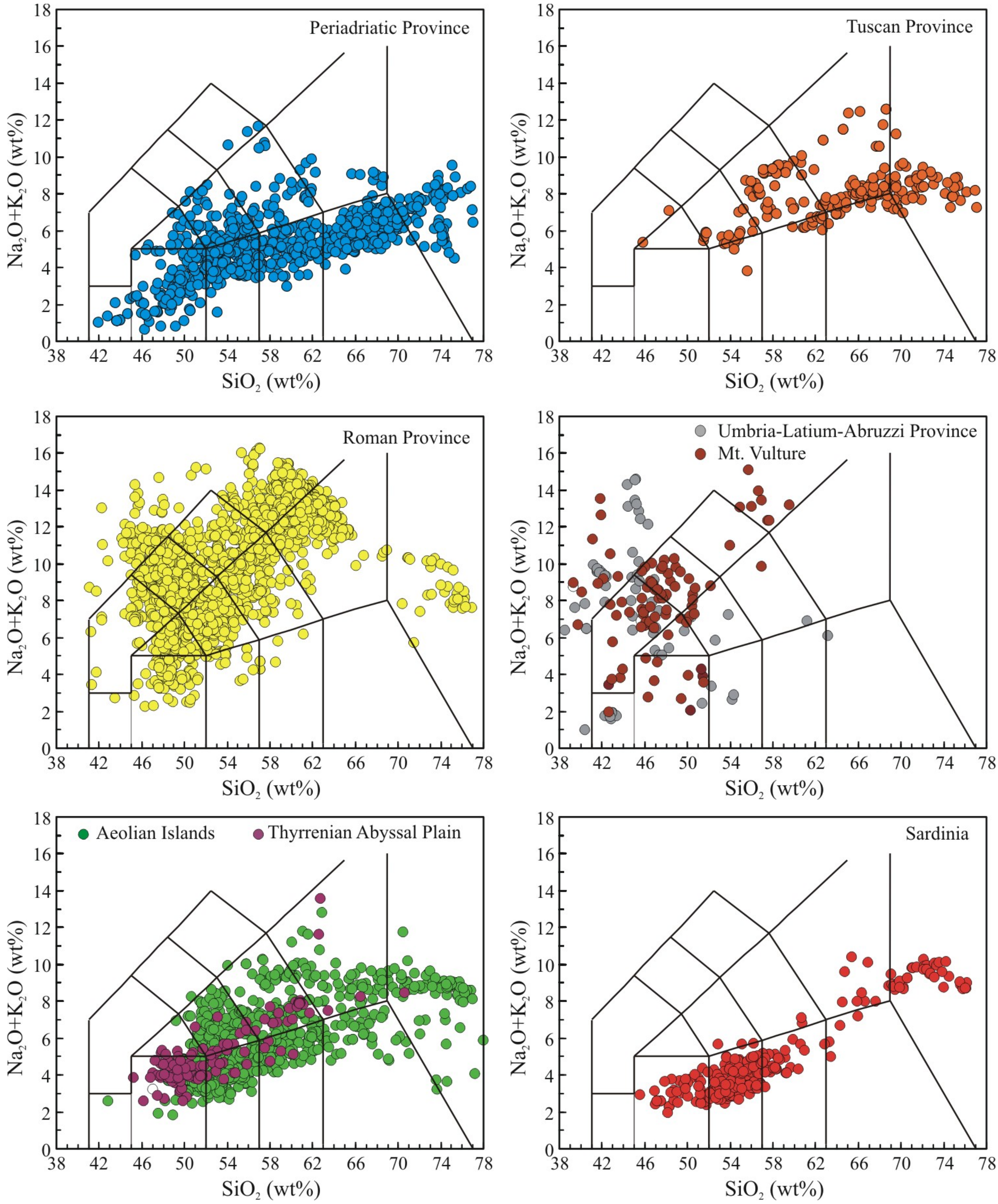
Apatite fission track data from the central-western Alps show that erosion and exhumation concentrated in the internal sectors of the belt and in the Lepontine dome in Late Oligocene to Early Miocene time, whereas, in the last 15-2 Ma exhumation migrated to more external positions (i.e., northward), in the area of the external crystalline massifs (Fügenschuh and Schmid, 2003; Garzanti and Malusà, 2008). On the contrary, erosion was more stationary in the Eastern Alps, where it created a narrower metamorphic belt (Rosenberg and Berger, 2009).

Figure 9. Alps and Apennines water divide

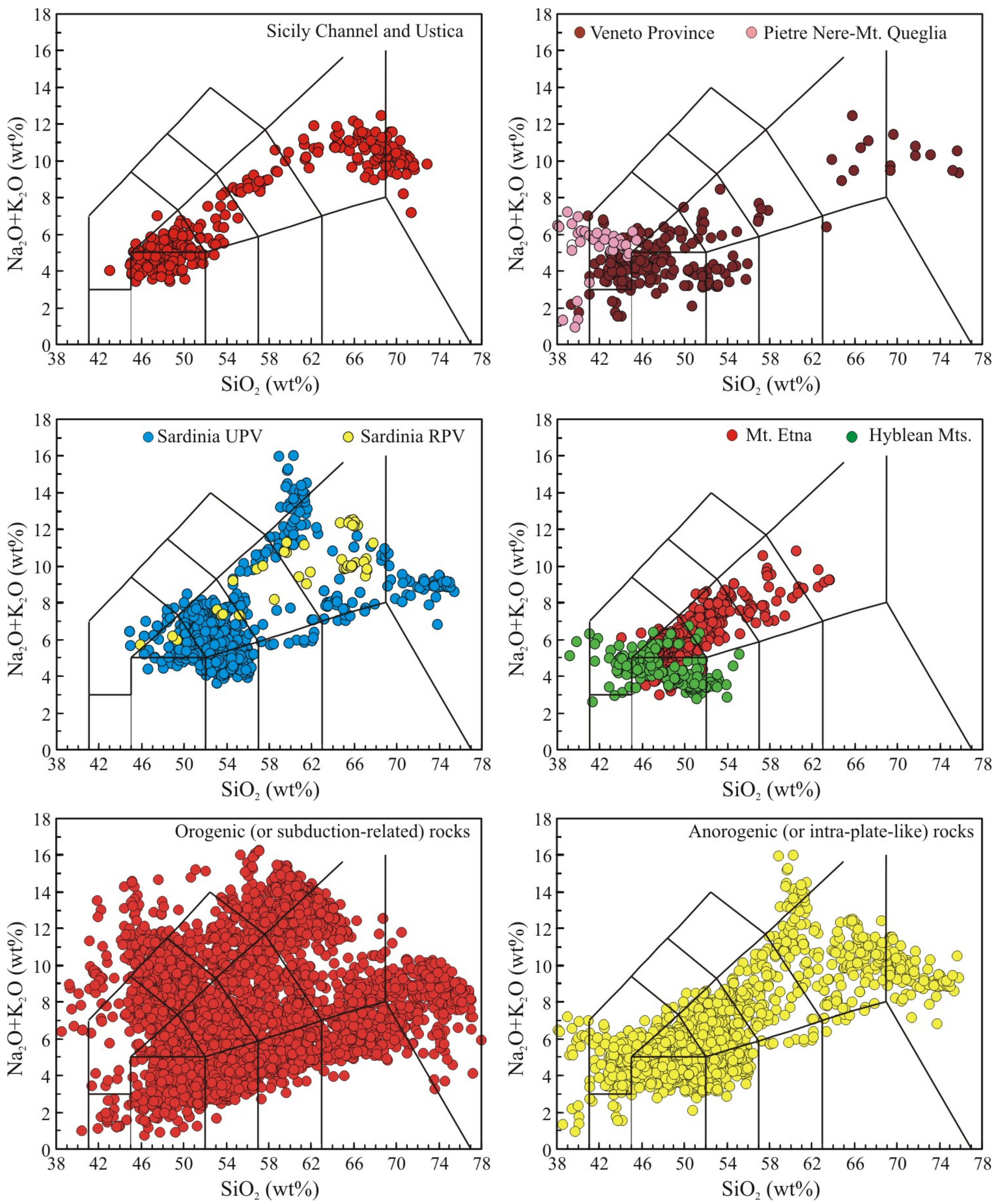


Alps and Apennines have a different relationship in the location and evolution of the water divide. In the Alps the water divide is more closely coincident with the highest peaks of the orogen. In the Apennines the water divide and the highest peaks are frequently shifted apart. The highest peaks often located to the east, apart for those zones where the subduction retreat is practically stopped and/or where the lithological contrast dominates the erosion rate (e.g., the Southern Apennines).

Figure 10. Total Alkali vs. Silica (TAS) diagram (Le Bas *et al.*, 1986) for the Cenozoic igneous rocks of Italy.

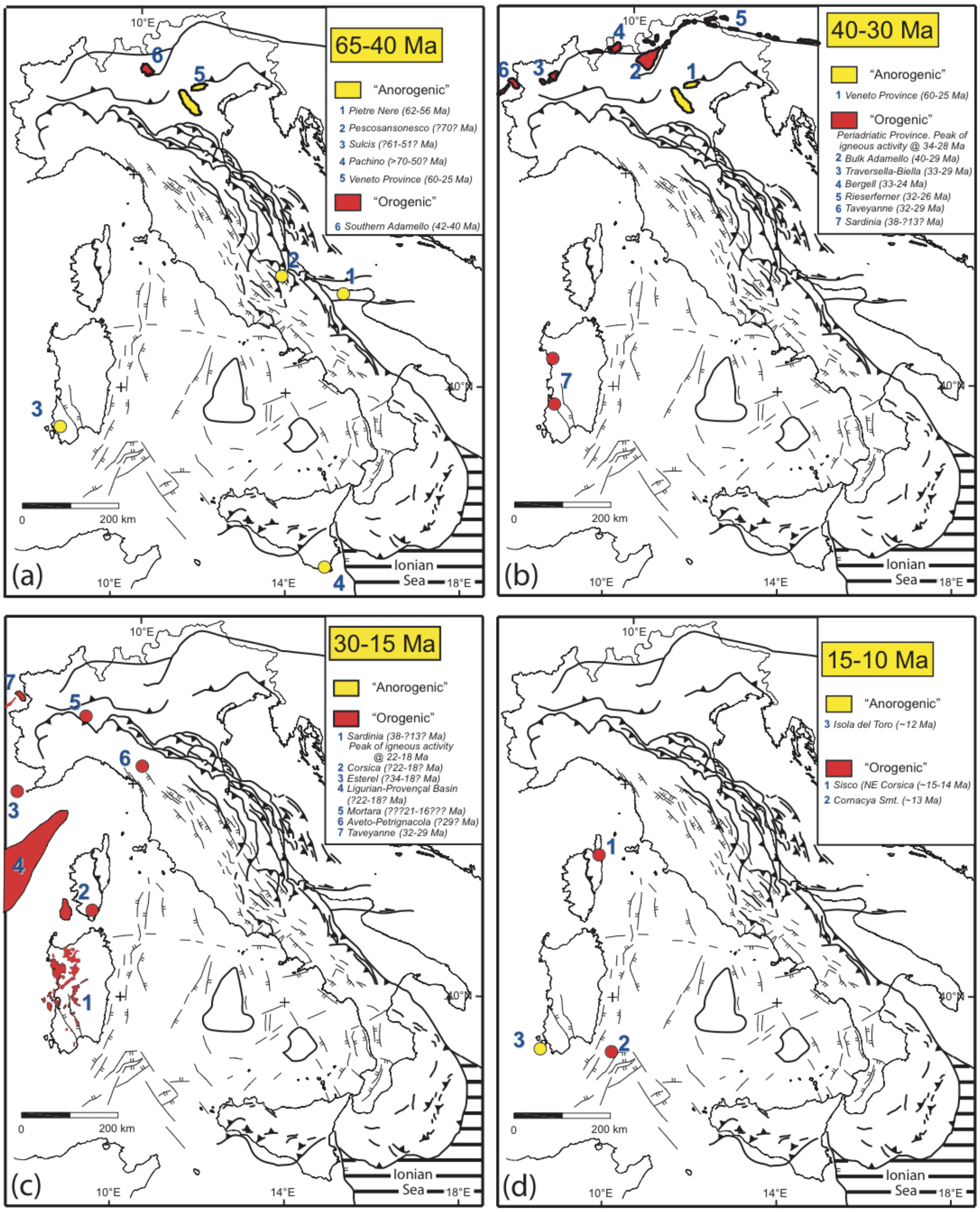


Top (a) = subduction-related (or “orogenic”) rocks. References in Lustrino *et al.* (2011).

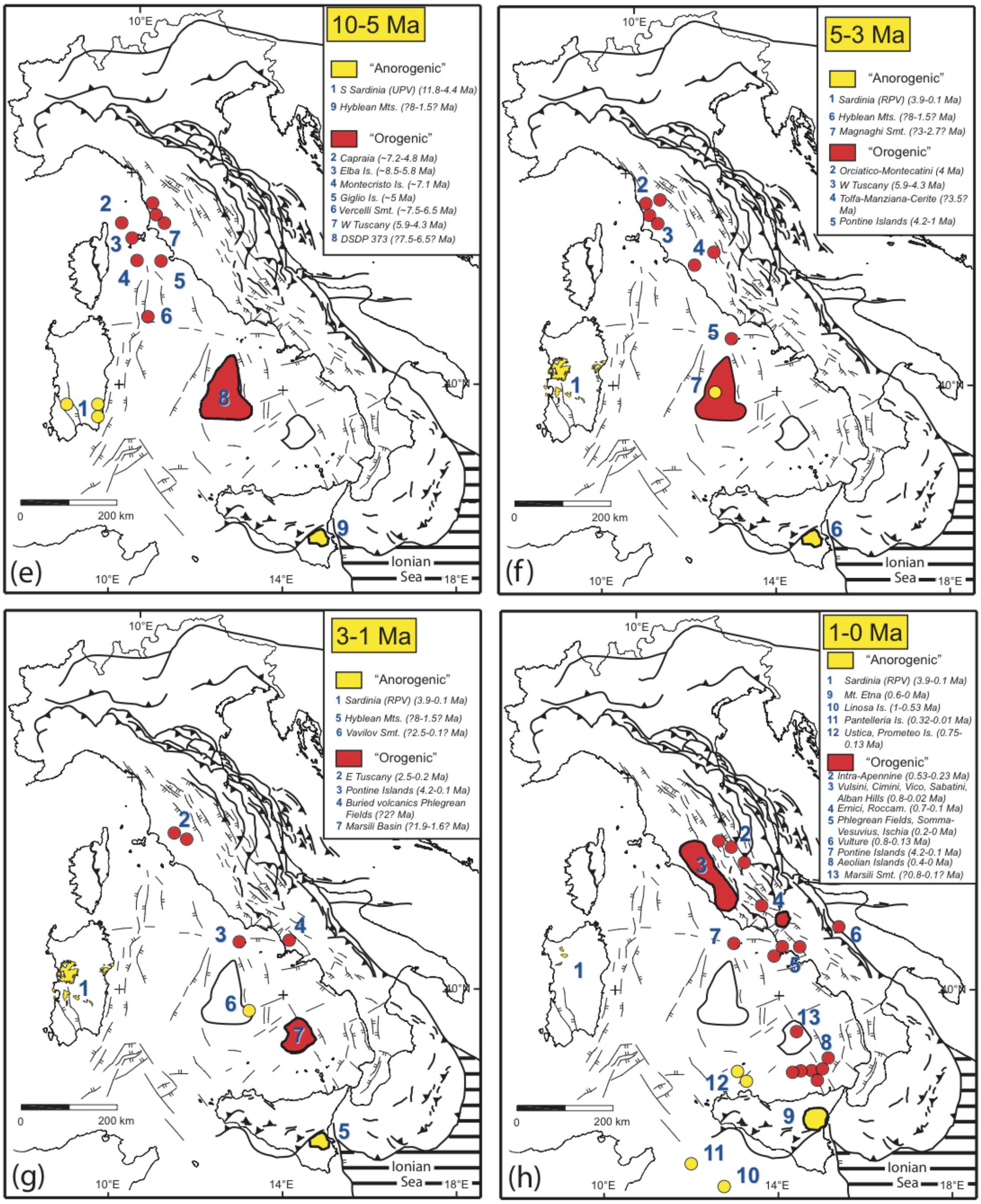


Bottom (b) = intra-plate-like (or “anorogenic) rocks. References in Lustrino and Wilson (2007) plus updates downloadable from the GEOROC web site (http://www2.mpch-mainz.mpg.de/~geo/Databases/GEOROC/Expert_Datasets.htm). The last two plots of fig. 7b show the Italian Cenozoic igneous rocks divided into “orogenic” and “anorogenic” groups. The full list of analyses (8754 whole rocks) and references can be requested to the second author.

Figure 11. Magmatic activity of the Italian area at different time spans, from 65 Ma to Present (a-h).



Yellow symbols: igneous rocks with chemical composition resembling magmas emplaced along active subduction zones (subduction-related or "orogenic"). Red symbols: igneous rocks with chemical composition resembling magmas emplaced in within-plate oceanic settings (intra-plate-like or "anorogenic").



For a better definition of "orogenic" and "anorogenic" terms used in geochemistry applied to igneous petrology the reader is referred to Wilson and Bianchini (1999), Lustrino and Wilson (2007) and Lustrino *et al.* (2011).

The most ancient Cenozoic igneous activity observable close to the Alpine Chain started during Middle Paleocene and continued intermittently until Late Oligocene (~65-25 Ma) in the Veneto region, eastern Po Plain. The igneous rocks of the Veneto region cover an area of ~2000 km^2 and are concentrated in four main volcanic districts (Lessini Mts., Marostica, Berici and Euganean Hills) with eruption of predominantly basic alkaline sodic lavas (mostly alkali basalts, hawaiites, benmoreites, trachytes) plus rarer Si-richer types (quartz trachytes and rhyolites; Fig. 10; references in Lustrino and Wilson, 2007).

Along the Alpine Chain, igneous activity developed mostly along the Insubric Line from ~42 to ~24 Ma (Fig. 11a-b-c), partially coeval with the igneous activity in the Veneto region (Rosenberg, 2004, and references therein). The bulk of the magmatism (>95% of the present-day outcrops) is emplaced in the time span 33-29 Ma and is represented by a syn-tectonic TTG (Tonalite-Trondhjemite-Granodiorite) association (Fig. 10). Minor gabbroic and dioritic stocks and even rarer calcalkaline and more exotic (i.e., lamproitic and kamafugitic) volcanic rocks are found mostly as dikes, especially in NW Alps (e.g., Venturelli *et al.*, 1984; Conticelli *et al.*, 2009, and references therein). With the exception of the plutonic rocks of southern Adamello, all the plutonic rocks emplaced along the Alpine Chain postdate or are coeval with the Eocene metamorphic peak (e.g., Rosenberg, 2004; Rosenbaum and Lister, 2005, and references therein).

Apennines and Sicily

The Apennines describe an arc from northwest Italy, down throughout the Italian peninsula, continuing to the southwest into Sicily and merging into the mountain belts (Maghrebides) of northwest Africa (Tunisia, Algeria and Morocco). The Apennine belt is mostly characterized by thin-skinned tectonics (Fig. 7; e.g., Bally *et al.*, 1986; Scrocca *et al.*, 2005, and references therein). The tectonic piling is radial, varying with continuity from N-NE (Northern Apennines) to NE (Central Apennines), E (Southern Apennines), E-SE (southernmost Apennines, Calabria) and S (Sicily; Fig. 3). The depth of the basal décollement plane at the front of the Apennines accretionary wedge varies significantly moving along strike (3 km in the Ionian Sea and 10 km in the Northern Apennines; Bigi *et al.*, 2003; Lenci *et al.*, 2004).

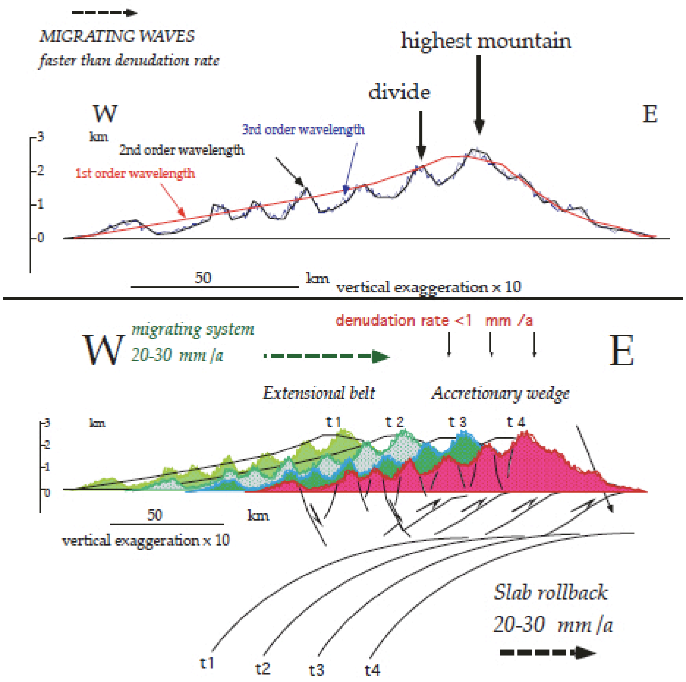
With the exception of the Calabro-Peloritani arc, the Apennines are mainly made up of Jurassic-Tertiary limestones (deposited both in shallow and deep-sea environments) and dolostones of African (Adriatic) origin (e.g., Patacca *et al.*, 1989; Dercourt *et al.*, 1992; Santantonio, 1994; Catalano *et al.*, 2001; Ciarapica and Passeri, 2005; Santantonio and Carminati, 2010), Late Tertiary foredeep silicoclastic deposits (Ricci Lucchi, 1986, 1990; Mutti *et al.*, 1999; Milli *et al.*, 2007), Early Miocene carbonate successions (e.g., Carannante *et al.*, 1988; Pescatore *et al.*, 1999; Brandano *et al.*, 2010) and subordinate ophiolitic melange (e.g., Montanini *et al.*, 2008; Molli, 2008; Tortorici *et al.*, 2008). The ocean, from which the ophiolitic bodies originated, separated the western margin of Adria from the southern European paleo-continental margin.

The geometry of the chain is shown by several seismic reflection and refraction studies acquired through the belt (Pieri, 1983; Scarascia *et al.*, 1994; Cernobori *et al.*, 1996; Barchi *et al.*, 1998; Improta *et al.*, 2000) and its foreland (e.g., Pieri and Groppi, 1981; Bally *et al.*, 1986; Sella *et al.*, 1988; Scrocca *et al.*, 2003). As shown in Fig. 5, the Adriatic lithosphere is characterized by average continental lithosphere thickness (~70 km) increasing towards the belt (~110 km; Calcagnile and Panza, 1981). The Ionian lithosphere is ~90 km thick (Panza *et al.*, 1992) and is rather interpreted as true oceanic (e.g., Finetti *et al.*, 1996; Catalano *et al.*, 2001) or transitional-to-continental (e.g., Calcagnile *et al.*, 1982; Cernobori *et al.*, 1996). A clearly defined Benioff-Wadati zone is present in Calabria-Peloritani-SE Tyrrhenian Sea, with hypocentre depths increasing towards NW (Finetti, 1982; De Voogd *et al.*, 1992; Catalano *et al.*, 2001; Chiarabba *et al.*, 2008).

Although the Apennines were deformed mainly during Neogene-Quaternary (e.g., Boccaletti *et al.*, 1990), remnants (thrust slices) of previously (Cretaceous-Paleogene) deformed rocks occur in the internal (western) portions of the belt (e.g., Abbate and Sagri, 1970; Principi and Treves, 1984). Associated with the east-directed tectonic piling, an “eastward” younging of foredeep sediments involved in the fold-and-thrust belt is recorded (e.g., Ricci Lucchi, 1986). Contemporaneous to compressional tectonics at the front, an “eastward” wave of extensional tectonics disrupted the inner portion of the belt, driving to the development of the Tyrrhenian Basin, of inland basins of western Italy and intra-mountain basins (e.g., Sartori, 1989; Bartole *et al.*, 1995; Barchi *et al.*, 1998; Cavinato *et al.*, 2002).

Although the Apennines are generally made of unmetamorphosed cover rocks, metamorphic rocks are found in the northern (Tuscany) and southernmost (Calabria, as discussed later) sectors of the chain (e.g., Vignaroli *et al.*, 2009 and references therein). Bimodal PT paths (HP metamorphism superimposed by a HT imprint) have been reconstructed for the rocks of the Tuscan Apennines and islands (e.g., Giglio and Gorgona Islands and Argentario Promontory; Apuane Alps; Rossetti *et al.*, 1999, 2001a). The age of the HP stage has been constrained to ~31-27 Ma (Brunet *et al.*, 2000) whereas the HT event occurred from 15 to 8 Ma (Kligfield *et al.*, 1986; Molli *et al.*, 2000).

Figure 12. Lateral migration of topography in the Apennines.

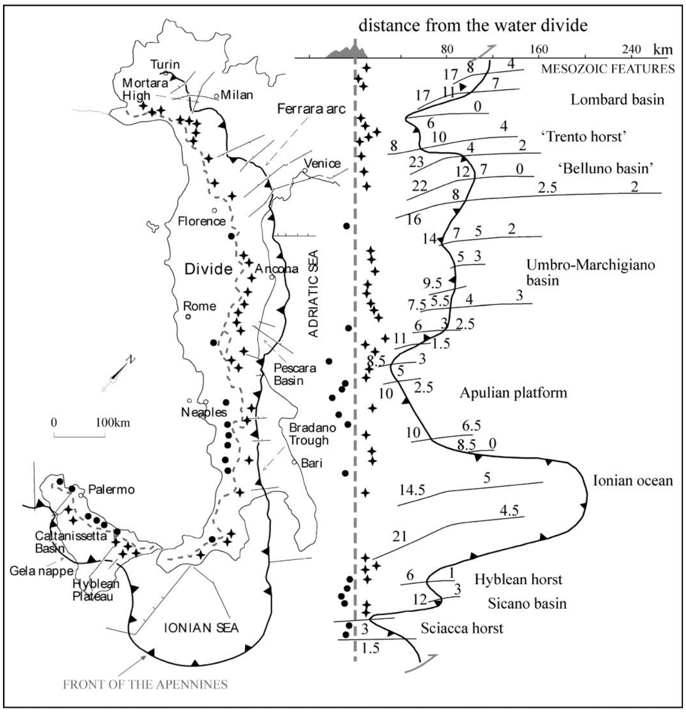


Above, idealised topographic section of the Apennines. The offset between divide and highest mountains appears to be associated to the faster effect of the tectonic ‘eastward’ migration with respect to the denudation rate. Deviations from this profile rely on strong lithological differences determining variable denudation rates, or slowing of the slab retreat rate of the Apennines subduction zone. At least three orders of topography can be distinguished: the first (> 100 km) due to geodynamic scale processes (subduction, depth of the decollement plane determining the volume of the accreted rocks in the prism, asthenospheric wedge under the belt); the second (5-15 km) determined by horsts and grabens spacing, or pre-existing folds; the third (0.5-2 km) related to smaller scale tectonics, lithologies and denudation rate. Below, interpreted relation between slab rollback and ‘eastward’ migration of the topography, in conjunction with the migration of the accretionary prism and the extensional belt. This simple model varies when slab rollback decreases, lithologies of the foredeep prevail in the emerged extensional belt decreasing frontal topography, and variations in depth of the frontal decollement plane (after Salustri Galli *et al.*, 2002).

Frequently, the highest mountain peaks do not coincide with the water divide (Salustri Galli *et al.*, 2002), but they are rather located toward the foreland to the East (Fig. 9). Since erosion is slower (0.7 mm/yr) than tectonics (1-10 mm/yr), the eastward propagation of the tectonic wave cannot be entirely compensated (Fig. 12). This generated a shift between the highest mountains and the water divide that remained more internal. In the Alps the lateral growth is not polarized as in the Apennines, and the water divide more frequently coincides with the highest peaks (Fig. 9). The salients and recesses in the Apennines accretionary prism front generally correspond to steeper or shallower foreland regional monocline respectively. These along-strike undulations appear controlled by the inherited grain of the Mesozoic Tethyan rift-related horsts and grabens, and still control the location of the water divide and the highest peaks (Fig. 13). The forward propagation of thrusts manly followed an in-sequence trend; however, an internal out-of-sequence thrusting can be inferred at the base of the foothills (Fig. 14).

The oldest Cenozoic igneous rocks cropping out along the Italian peninsula are found in southern Italy at Punta delle Pietre Nere, Gargano, Apulia region (Fig. 11a). This is a Paleocene (~62-56 Ma) small outcrop (~800 m^2) of intrusive mafic alkaline rocks (mela-syenite and melagabbro) with lamprophyric (i.e., K-rich) affinity (Vollmer, 1976; de Fino *et al.*, 1981; Bigazzi *et al.*, 1996; Conticelli *et al.*, 2009; Fig. 11a). A very small dyke swarm, about 15 m wide, containing 1-6 m thick dykes has been found in central Italy (Monte Queglia, Abruzzo region; Fig. 11a). These dykes, whose age has been constrained to Eocene time (<54 Ma), show overall mineralogical and chemical compositions resembling the Punta delle Pietre Nere rocks (Barbieri and Ferrini, 1984; Durazzo *et al.*, 1984; Vichi *et al.*, 1995; Fig. 10a). From Late Paleocene to Late Miocene no igneous activity is recorded along peninsular Italy. The only exception are Oligocene (~29 Ma) conglomerates, up to 400 m thick, rich in volcanic fraction (with boulders up to 1-2 m^3) found in the Aveto-Petrignacola formation (Northern Apennines; Mattioli *et al.*, 2002a; Fig. 11c) and a ~30 Ma-old buried volcano, covered by thousands of meters of sediments in the Po Valley (Mortara volcano; Mattioli *et al.*, 2002b; Fig. 11c). The chemical and mineralogical composition of the Aveto-Petrignacola and Mortara rocks is quite similar, with abundance of andesitic, dacitic and rhyolitic composition with calcalkaline affinity (Fig. 10b). These rocks share mineralogical, petrographic and geochemical similarities also with the coeval (~32 Ma-old) volcaniclastic sandstones of the Taveyanne formation (NW Alps; Ruffini *et al.*, 1997; Fig. 11c).

Figure 13. Relation between the regional monocline, the accretionary prism front, the divide and the highest topography.



Relation between the regional monocline around the Apennines (grey lines; the numbers indicate the monocline dip in degrees), the accretionary prism front (barbed line), the divide (dashed line; modified after Mariotti and Doglioni, 2000) and the highest topography (dots and crosses) in a map view (left), and normalized to the divide (right), marked with dots. In the central-northern Apennines the highest mountains are more displaced toward the foreland than along the southern Apennines where the thick continental lithosphere of the Apulian Platform encroached the trench and slowed the subduction zone. Salients and recesses of the accretionary prism follow the inherited Mesozoic grain (i.e., horsts and grabens). The water divide and the highest peaks miscorrelation appears to be controller as well by the Mesozoic architecture. The absence of a large offset in Calabria could be attributed to salt in the decollement at the base of the accretionary prism in the Ionian, to the high denudation rate of the crystalline rocks inland, and the presence of relevant transfer zones such as the Crati and Catanzaro grabens (after Salustri Galli *et al.*, 2002).

Other sparse outcrops of Oligocene-Miocene igneous rocks are represented by pelitic to arenitic levels rich in volcanic fragments: the Bisciaro Formation ("Livello Raffaello", possibly Aquitanian, central Italy; Montanari *et al.*, 1991), the Upper Oligocene-Aquitanian marls of the Tertiary Ligurian-Piemontese Basin (e.g., "Marne di Rigoroso", Monferrato; Ruffini and Cadoppi, 1994; "Marne di Antognola"; Morandi *et al.*, 1991), the Late Oligocene volcaniclastic levels of the Ranzano Basin (Northern Apennines; Cibin *et al.*, 1998), the Late Oligocene-Early Miocene Calabro-Lucano Flysch (Southern Apennines; Critelli and Monaco, 1993), the Late Oligocene volcaniclastic turbidites of the Tufiti di Tusa Formation (Northern Sicily; Carbone *et al.*, 2005).

The igneous activity resumed from latest Miocene until present days. From ~8.5 to ~0.2 Ma a discrete igneous activity generated series of volcanoes and plutonic stocks mostly in the Tuscany region (North-Western Apennines) and offshore northern Tyrrhenian Sea (Poli, 2003; Peccerillo, 2005; Peccerillo and Lustrino, 2005; Avanzinelli *et al.*, 2009; Fig. 11e). These rocks are grouped under the name of Tuscan Magmatic Province, and are represented both by crust- and mantle-derived melts, i.e., with pure crustal anatectic rhyolitic melts as well as liquids of pure mantle and hybrid (crust + mantle) origin (Fig. 10b). Plagioclase- and leucite-free lamproitic rocks are found in association with shoshonites, and high-K calc-alkaline rocks. Despite the name of the province, the southern outcrops of the Tuscan magmatic province crop out in northern Latium region (Tolfa-Cerite districts; Peccerillo, 2005; Fig. 11f).

Italy is particularly famous among Earth scientists for its Plio-Quaternary igneous activity developed along the Tyrrhenian coast with exotic chemical and mineralogical composition (Peccerillo, 1985, 2005; Beccaluva *et al.*, 1991; Conticelli *et al.*, 2002; Peccerillo and Lustrino, 2005; Avanzinelli *et al.*, 2009; Lustrino *et al.*, 2011). These became the subject of detailed mineralogical, petrographic and geochemical investigations more than one century ago. Washington (1906) recognised and defined for the first time the Roman Magmatic Province on the basis of the common presence of abundant leucite and the strong degree of SiO_2-undersaturation. Since then, the volcanoes of this region have been fully investigated, and a wealth of geochemical, mineralogical, petrographic, volcanological and geophysical data is now available. The igneous activity in the Roman Magmatic Province started during early Pliocene (~4.2 Ma) and is still active in the southernmost sectors (Mt. Vesuvius, Ischia Island and Phlegrean Fields; Peccerillo, 2005, and references therein; Fig. 11g-h). From North to South, the main volcanic edifices are Vulsini Mts., Vico Volcano, Sabatini Mts., Alban Hills, Ernici Mts., Mt. Roccamonfina, Pontine Islands, Procida, Vivara and Ischia Islands, Somma-Vesuvius and the Phlegrean Fields (Fig. 11g-h). The oldest products (~4.2-1 Ma) crop out in the Pontine Island district as acid volcanic rocks (mostly trachytes and rhyolites; Fig. 10b). The SiO_2-oversaturated composition of these rocks strongly contrast with the SiO_2-saturated to strongly SiO_2-undersaturated compositions of the Roman Magmatic Province products. On the basis of this difference, the Pontine Islands rocks are grouped in the Tuscan anatectic province by several authors (Peccerillo, 2005; Avanzinelli *et al.*, 2009, and references therein) but are grouped in Figure 10b together with the Roman Magmatic Province. With the exception of the Pontine rocks and the thick volcanic successions buried beneath the Campania plain (where more than 1800 m thick andesites have been drilled by the national oil company), the main rock types belong to potassic SiO_2-saturated (e.g., shoshonitic) and strongly potassic SiO_2-undersaturated series (e.g., leucitites, tephrites to phonolites; Peccerillo, 2005; Conticelli *et al.*, 2009; Avanzinelli *et al.*, 2009; Lustrino *et al.*, 2011, and references therein).

A completely different igneous activity (in terms of chemical composition of the volcanic rocks) developed in the central sectors of Apennines in a restricted time range. The rocks that belong to this activity are grouped in the Umbria-Latium-Abruzzo ultra-alkaline (or Intra-Apennine) Province (~0.6.0.4 Ma; Stoppa and Lavecchia, 1992; Castorina *et al.*, 2000; Stoppa *et al.*, 2003; Fig. 11h). The peculiar feature of this province is the presence of numerous monogenetic volcanoes, located along main extensional faults, with strongly SiO_2-undersaturated composition, exotic mineralogical paragenesis and very limited areal extension. The main rock types are strongly SiO_2-undersaturated melts like melilitolites, melilitites, Ca-carbonatites and ultrapotassic rocks (kamafugite group, with coppaelite and venanzite type localities near Cupaello and San Venanzo villages; Fig. 10b).

In terms of geographic position (southern Apennines, close to the Apennine thrust front), mineralogy (e.g., presence of exotic minerals like haüyne) and chemical whole-rock composition, the Mt. Vulture (or Lucanian Province) can be considered isolated from the main Roman Magmatic Province (Fig. 11h). The igneous activity developed during Pleistocene (~0.8-0.1 Ma; i.a., Melluso *et al.*, 1996; Beccaluva *et al.*, 2002; Peccerillo, 2005; De Astis *et al.*, 2006) with emission of pyroclastic rocks with minor lava flows and domes. The main rock types of Mt. Vulture are basanites and trachy-phonolites with minor foidites, tephrites, phono-tephrites and melilitites (Fig. 10b), less potassic than the Roman district rocks and with haüyne joining leucite as the main feldspatoid. A carbonatitic (alvikite) small lava flow has been recently discovered in this district (D’Orazio *et al.*, 2007, 2008).

In North-East Sicily, one of the largest volcanoes of the entire Mediterranean area crops out: Mt. Etna (Fig. 11h). The first igneous activity of this volcano has been dated ~0.6 Ma (Gillot *et al.*, 1994) and since then it is still active. The first products are tholeiitic to transitional lavas, later evolved towards more sodic alkaline magma compositions (Tanguy *et al.*, 1997; Schiano *et al.*, 2001; Armienti *et al.*, 2004; Corsaro and Pompilio, 2004; Lustrino and Wilson, 2007; Viccaro and Cristofolini, 2008; Fig. 10a). The most recent products have a mildly alkaline affinity and relatively evolved compositions (hawaiites, mugearites, benmoreites, trachytes; Fig. 10a).

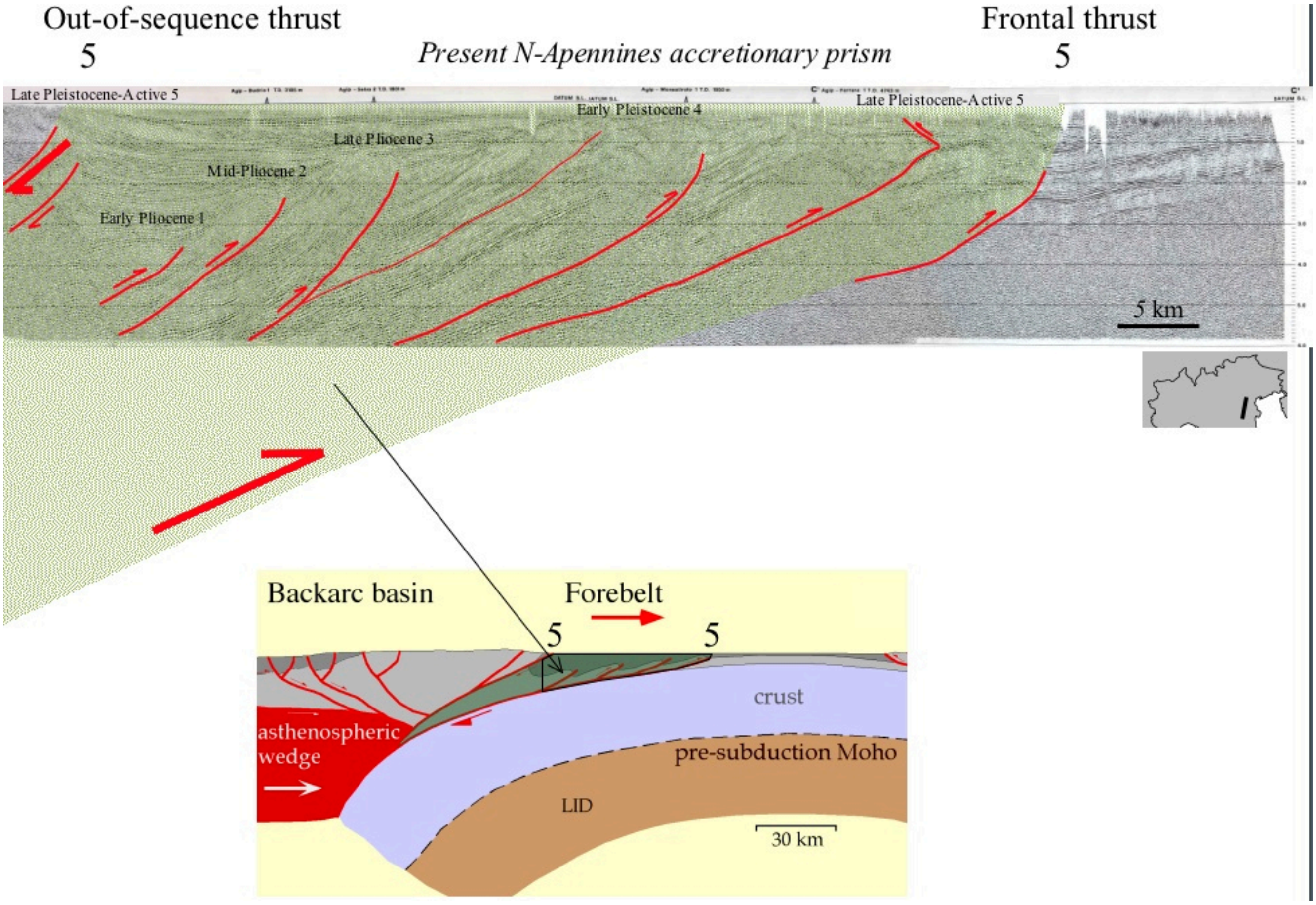
Few tens of km south of Mt. Etna, the Hyblean Plateau volcanic province crops out (Fig. 11a-e-f-g). The most recent igneous activity (Miocene-Pliocene) includes abundant tholeiitic basalts, plus minor alkaline sodic mafic rocks basanites, alkali basalts, hawaiites, nephelinites and ankaratrites (Di Grande *et al.*, 2002; Peccerillo, 2005, and references therein; Fig. 10a).

Po Plain and Adriatic Sea

The Po plain is the foreland basin of northern Apennines and Southern Alps (e.g., Doglioni, 1993). Active thrusts, related to tectonics in the Apennines (e.g., Pieri, 1993; Scrocca *et al.*, 2007; Carminati *et al.*, 2010), are buried beneath the Quaternary sediments of the southern Po Plain (Fig. 14 and 15). The inactive front of the central Southern Alps (i.e., west of the Giudicarie Line) is also buried beneath Po Plain sediments in the so-called “Milano Belt” (Laubscher, 1990). The south-verging front of the Southern Alps and the north-verging front of the Northern Apennines merge south of Milano, as shown by seismic lines (Pieri and Groppi, 1981; Consiglio Nazionale delle Ricerche, 1992; Fig. 3). The Pliocene-Quaternary sedimentary pile, made up of shallow-sea to continental mainly siliciclastic rocks (e.g., Amorosi, 1996), is as thick as 8 km, implying fast subsidence rates (>1 mm/yr) during the last 5 Ma (Fig. 8 and 15).

Tectonic (e.g., lithospheric flexure, faulting and folding) and sedimentological (e.g., compaction) processes control these long-term vertical velocities (Carminati *et al.*, 2003a). Shorter-term subsidence pulses in the area are controlled by ice cycles (e.g., Carminati *et al.*, 2003b) and by human activities (Carminati and Martinelli, 2002). Geodynamics and climatological changes are the main drivers of the sedimentation (Vittori and Ventura, 1995; Garzanti *et al.*, 2002), which has filled the basin with alternate layers of sand and clay. Gravel is present only near the Apennines chain in alluvial fan deposits (Ori, 1993; Amorosi et. al., 1996), and in fans in the foothills between the Alps and the Po River extending from the outlets of the Alpine rivers into the Plain (Guzzetti *et al.*, 1997).

Figure 14. Timing of thrusting at the front of the Apennines accretionary prism



The front of the Apennines accretionary prism buried beneath the Po Basin shows an in-sequence foreland thrust propagation (1-5) since the Late Miocene-Early Pliocene up to Present. However, at the southern margin, an active out-of-sequence thrust deforms Peistocene sediments (5). The accretionary prism off-scraps only the shallower part (dominantly sedimentary cover) of the subducting lithosphere. Most the foredeep basin is located on top of the accretionary prism (piggy-back basin). In the section below the geodynamic interpretation. Seismic section after Pieri (1983).

The Adriatic Sea is the strongly segmented foreland basin of both Apennines and Dinarides-Hellenides (e.g., Royden *et al.*, 1997; Carminati *et al.*, 2004b; Scrocca, 2007). The Adriatic Italian offshore is characterized by fast subsidence rates (up to 1 mm/yr) in the northern part (e.g., Carminati and Di Donato, 1999), and rather by tectonic uplift (active since the Middle Pleistocene with rates of 0.3-0.5 mm/yr) in the southern Apulian part (Doglioni *et al.*, 1994). The sedimentary fill of the basin is dominated by turbiditic siliciclastic sediments. As shown in Fig. 3, the front of the Apennines is located in the western portion of the northern Adriatic Basin and reaches the middle of the central Adriatic Basin (Scrocca *et al.*, 2007). These fronts are buried under Pliocene-Recent sediments. The southern Adriatic basin, as well as onshore Apulia, form the foreland area of the Apennines, the Apenninic front being positioned well inland along the Bradanic Trough (e.g., Doglioni *et al.*, 1994; Fig. 3).

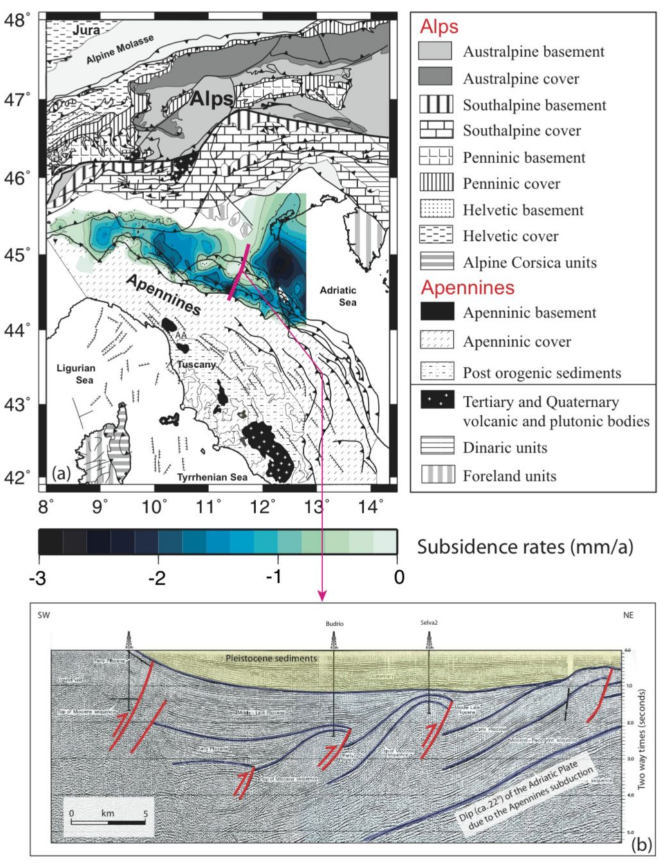
Sardinia

The geological structure of Sardinia is the result of several tectono-magmatic processes that begun with the Late Paleozoic compressional tectonics that led to the consumption of large oceanic masses (Paleo-Tethys) and the amalgamation of continental plates into the super-continent Pangea (tectono-magmatic phase locally ascribed to the Hercynian Orogeny; see Compagnoni et al., this volume). The back-bone of this island is made up of plutonic and metamorphic rocks related to the shortening of the Hercynian Orogeny and the following relaxation tectonics that led to a disruption of the recently amalgamated super-continent. During the Carboniferous-Permian, Sardinia (and Corsica) experienced abundant production of calcalkaline to high-K calcalkaline granitic-granodioritic magmatism (e.g., Tommasini *et al.*, 1995; Ferré and Leake, 2001) plus minor gabbroic-tonalitic and volcanic activity (calcalkaline to sodic alkaline dykes), partially coeval with, or pre-dating continental rifting that dismembered Pangea (see Lustrino, 2000, for a review).

Up to the Eocene, Sardinia and Corsica were part of the southern margin of Europe. During Late Eocene, a first extensional phase, roughly coeval with extensional tectonics in central Europe (Rhine Graben) and SE France (Alés, Manosque-Forcalquier and Camargue Basins; Séranne, 1999) developed in SW Sardinia with the formation of the E-W-trending Cixerri Graben. A second, more important, Late Oligocene extensional phase affected the entire Sardinia-Corsica micro-plate and resulted in the development of NE-SW directed rifting system that evolved into continental drifting during Late Oligocene-Early Miocene (e.g., Alvarez *et al.* 1974; Vigliotti and Langenheim 1995; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007, and references therein). The Oligocene rifting in Sardinia was synchronous to the Ligurian-Provençal basin rifting, dated from the Oligocene (~34 Ma) to the middle Aquitanian (21 Ma) according to the age of the sediments drilled in the Gulf of Lion, just beneath the break-up unconformity (Gorini *et al.*, 1993). Burdigalian extensional tectonics is also recorded in onshore grabens (Maerten and Séranne 1995). At a poorly constrained time (~22 Ma) the Sardinia-Corsica block started to rotate counter-clockwise away from the southern Europe paleo-margin, with a rotation pole located around the gulf of Genoa (Liguria, NW Italy; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007). The rifting and drifting tectonics associate with this rotation led to the formation of the Ligurian-Provençal Basin, possibly floored by oceanic crust (e.g., Gueguen *et al.*, 1998; Rollet *et al.*, 2002; Schettino and Turco, 2006). The rotation ended around the Langhian (~15-16 Ma; see Lustrino *et al.*, 2009, for a review) when intense igneous activity occurred (Brotzu, 1997; Lustrino *et al.*, 2004, 2009). The Oligocene-Aquitanian tectonics of Sardinia is dominated by N-S trending normal faults (present-day coordinates) bounding a graben system that cut through pre-Alpine basement Sardinia from North to South for a total length of about 220 km and an average width of 50 km (Casula *et al.* 2001; Faccenna *et al.*, 2002; Cherchi *et al.*, 2008). The Sardinian Trough (also known as *Fossa Sarda*) is filled by up to 2 km of Upper Oligocene syn-rift (mainly non-marine sediments followed by fluvial-deltaic sediments, marls and silts) and Burdigalian-Messinian post-rift (marine marl-silt successions) deposits (Cherchi *et al.*, 2008, and references therein). Starting from the Pliocene a new, NW-SE trending graben developed in SW Sardinia (Campidano graben; e.g., Balia *et al.*, 1991), partially overlapping the southern sector of the *Fossa Sarda*.

As it concerns the Cenozoic igneous activity of Sardinia, the oldest products are Paleogene (~61-51 Ma) lamprophyric dykes (camptonite) drilled in the SW sector of the island (Sulcis area; Maccioni and Marchi, 1994; Fig. 11a). These dykes show major and trace element content roughly similar to the Pietre Nere and Mt. Queglia alkaline rocks (see above; Fig. 11a). Apart for these volumetrically insignificant products, the bulk of the igneous activity started during Late Eocene. With the exception of a single sub-volcanic outcrop dated at ~38 Ma (Lustrino *et al.*, 2009; Fig. 11b), the bulk of the igneous activity in Sardinia started at ~32 Ma, during the peak of the igneous activity along the Insubric Line. The peak of the activity was concentrated around 22-18 Ma, in concomitance with the fastest counter-clockwise rotation of the island (Lustrino *et al.*, 2004, 2009; Gattacceca *et al.*, 2007; Fig. 11b-c). During the peak of the volcanic activity in Sardinia, the magma production along the Alpine Chain and in the Veneto area had already ceased. The igneous rocks of Sardinia are volcanic (<1% of plutonic rocks), essentially in ignimbritic facies, with minor lava flows and domes. These have mostly medium- to high-K calcalcaline composition, with rare arc-tholeiitic types. Very rare peralkaline compositions are produced during the last stages of igneous activity (with the type-locality of comendites in S. Pietro Island, Sulcis district, SW Sardinia; Morra *et al.*, 1994). Dacitic to rhyolitic ignimbrites are the most abundant rock types, followed by andesitic, basaltic andesitic and minor basalts (Morra *et al.*, 1997; Lustrino *et al.*, 2004, 2009, and references therein; Fig. 10b). The igneous activity in Sardinia stopped around 15-13 Ma, after the end of counter-clockwise rotation of the Sardinia-Corsica micro-block (Beccaluva *et al.*, 1985; Speranza, 2002; Gattacceca *et al.*, 2007; Lustrino *et al.*, 2009).

Figure 15. Long term subsidence rates for the Po Plain.



(a) Long term (averaged over the last 1.4 Ma) subsidence rates for the Po Plain (after Carminati and Di Donato, 1999). AA: Apuane Alps. (b) Seismic section through the front of the Northern Apennines (interpreted from Pieri, 1983). The trace of the section is shown in panel a.

Early Oligocene-Early Miocene (~34-16 Ma) calcalkaline and high-K calcalkaline rocks (andesites, microdiorites, dacites and rhyolites, often emplaced as ignimbritic flows) are also found along the Provençal coast in southern France (Nice-Cap D'Ail area; e.g., Ivaldi *et al.*, 2003; Beccaluva *et al.*, 2005; Lustrino *et al.*, 2), SE Corsica (e.g., Ottaviani-Spella *et al.*, 2001) and offshore western Corsica (Rossi *et al.*, 1998). In the late Middle Miocene (~11.8 Ma) a new phase of volcanic activity developed in Sardinia, with mineralogical, chemical and isotopic characters completely different from the Late Eocene-Middle Miocene phase (Lustrino *et al.*, 2007b; Fig. 11d). The products of this second volcanic phase are low-volume, low degree-melting products with sodic alkaline compositions, ranging from basanitic to trachytic (Fig. 10b), found only in the southern sectors of Sardinia (Fig. 11d-e). The youngest products of this second phase of Cenozoic igneous activity of Sardinia (called RPV; Radiogenic Pb Volcanics; Lustrino *et al.*, 2000, 2007a) are dated 4.4 Ma. After ~0.6 Ma a new volcanic activity (third phase) developed in the central-northern sectors of Sardinia (Fig. 11f-g-h) with mildly sodic alkaline and tholeiitic compositions (Fig. 10a) characterized by very peculiar trace element and isotopic features compared to the rest of the entire circum-Mediterranean Cenozoic igneous rocks (see below). The last products of this activity (called UPV group, i.e., Unradiogenic Pb Volcanics) are as young as 0.1 Ma (Gasperini *et al.*, 2000; Lustrino *et al.*, 2000, 2007a; Fig. 11h).

Calabria and Sicily

After the cessation of the Sardinia-Corsica block rotation (~15 Ma), the eastward extensional wave shifted to the East of this block, leading to the extension and break-up of its easternmost margin: the Calabrian-Peloritani terrane (see the movie). This terrane was separated from Sardinia probably during the Langhian and drifted towards the SE up to its present position. The Calabrian-Peloritani Arc (Fig. 3) is the link between the Southern Apennines and the Maghrebian belt, but differs from these being constituted of crystalline basement nappes, with a top-to-the-East to top-to-the South tectonic polarity. These nappes are both of continental (Hercynian granitoids and meta-granitoids) and oceanic (Neo-Tethys) origin (Amodio Morelli *et al.*, 1976; Liberi *et al.*, 2006; Johnston and Mazzoli, 2009; Vignaroli *et al.*, 2009), plus sedimentary cover metamorphosed up to blue schist facies. HP-LT metamorphism in ophiolites and abyssal sediments from Calabria has been alternatively dated at 38–33 Ma (Rossetti *et al.*, 2001b) or ~30 Ma (Iannace *et al.*, 2007), thus possibly locating the HP event at the Eocene–Oligocene boundary (Vignaroli *et al.*, 2009). No igneous activity is recorded in Calabria during the Cenozoic. This accretionary prism was thrust onto the Apenninic units, a continental platform formed by low-grade Paleozoic crystalline basement rocks covered by Mesozoic carbonate deposits, during the Early Miocene (Amodio Morelli, 1976, Ietto and Barilaro, 1993).

The Peloritani Mts. Belt (north-eastern Sicily) is composed mostly of crystalline units, metamorphosed mainly during the Hercynian orogenesis (De Gregorio *et al.*, 2003), thrust onto the sediments of the African margin during the Tertiary. Late Eocene-Langhian syn-orogenic deposition, controlled by a combination of active thrusting, regional subsidence and sea-level change, was followed by sedimentation in basins controlled by normal faults from the Serravallian to the Pleistocene (Lentini *et al.*, 1995). The Recent evolution of the Peloritani Mts. Belt is characterized by progressive uplift of the southern margins of the Tyrrhenian Basin, and local active subsidence related to down-faulting (Lentini *et al.*, 1995).

Similarly to the Apennines, the bulk of the island of Sicily (south-west of the Taormina Line) is made up of Mesozoic-Cenozoic sediments thrust southward (although clockwise block rotations and oblique-slip thrusting also occurred) during the Neogene (Scandone *et al.*, 1974; Oldow *et al.*, 1990; Channell *et al.* 1990; Catalano *et al.*, 2000). The Iblean Plateau, mainly comprising Mesozoic to Lower-Middle Miocene carbonatic rocks, discontinuously overlain by more recent sedimentary and volcanic rocks, is the foreland area of the Apennines-Maghrebian belt of Sicily.

Tyrrhenian Sea

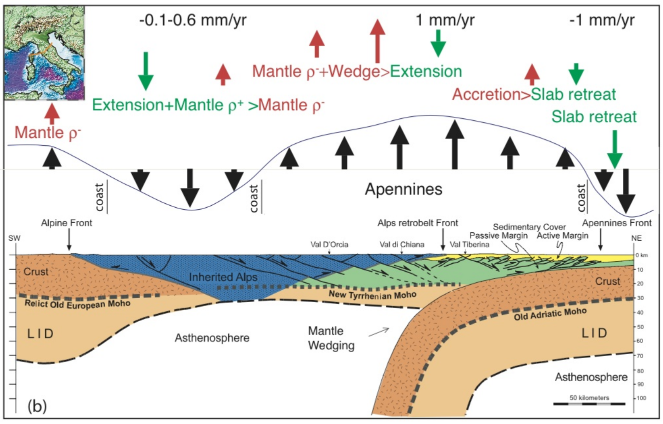
During the last stages of the Late Eocene-Middle Miocene igneous activity of Sardinia (~15 Ma), northern Corsica was characterized by the intrusion of a small dyke swarm in the Sisco area (NE-most Corsica; Fig. 11d), with a peculiar lamproitic (i.e., SiO_2- and MgO-rich, Al_2O_3- and CaO-poor ultrapotassic leucite-free) composition (e.g., Peccerillo *et al.*, 1988; Conticelli *et al.*, 2009). These are considered as the first magmatic activity associated with the opening of the Tyrrhenian Sea. Avanzinelli *et al.* (2009) grouped the Sisco lamproite with the Cornacya seamount (~12 Ma; SE Sardinia; Mascle *et al.*, 2001; Fig. 11d) and the oldest products of Capraia Island (~7.2 Ma; Tuscan archipelago; Chelazzi *et al.*, 2006; Gasparon *et al.*, 2008; Fig. 11e) in the Corsica Magmatic Province on geochemical and petrological grounds.

A new phase of rifting affected easternmost Sardinia, its offshore and the Tyrrhenian Sea starting from Middle Miocene. The eastern Sardinia continental margin is characterised by half grabens filled by syn-rift sediments of Late Tortonian to Messinian age, including thick Messinian evaporites, while the post-rift Pliocene-Quaternary sequence drapes the previous stratigraphic units. A pre-rift terrigenous sedimentary sequence of Serravallian-Tortonian age occurs on the uppermost Sardinia margin (Sartori *et al.* 2001). The basement of these basins is characterised by crystalline-metamorphic Variscan and Alpine rocks (including Tethyan ophiolites) that recorded compressive deformation (Sartori 1986; Bigi *et al.* 1990; Doglioni 1992; Guerrera *et al.* 1993), covered by syn-orogenic clastic sediments.

The Tyrrhenian Sea is a triangular extensional basin, more than 3600 m deep in its central part, crossed by a major (probably lithospheric scale) discontinuity known as the 41st Parallel Line (Savelli and Wezel, 1979; Spadini and Wezel, 1994; Bruno *et al.*, 2000). North of this lineament (in the Northern Tyrrhenian Basin), geophysical studies indicate the presence of a stretched continental crust ~20 km thick (Panza *et al.*, 1992). Here only small amounts of igneous rocks (Tuscany Province) have been found (e.g., Peccerillo, 2005, and references therein). South of the 41st Parallel Line, more intense stretching of the continental crust (the crust is <10 km thick; e.g., Morelli *et al.*, 1970) led to a diachronous oceanization in two sub-basins (Vavilov and Marsili, characterized by crust thinner than 6 km; e.g., Kastens *et al.* 1988; Sartori 1990). The depth of the Moho increases towards the continental margins reaching depths of 30 km along the Corsica-Sardinia margin. The lithosphere is as thin as 30 km under the central part of the basin (Panza *et al.*, 1992), where large heat flow values (>150 mW/m^2) are recorded (e.g., Della Vedova *et al.*, 1984; Zito *et al.*, 2003).

In the Tyrrhenian domain, similarly to Corsica and the Apuane Alps, extensional deformation is rlargely accommodated by ductile detachment faults disrupting an overthickened (likely Alpine) belt, since the Late Oligocene-Miocene (Carmignani and Kligfield, 1990; Jolivet *et al.*, 1990; Molli *et al.*, 2006; Fig. 16). The first subsidence episode was recorded by Langhian sediments deposited along the eastern coast of the Corsica island (Orszag-Sperber and Pilot, 1976), coeval with the Sisco igneous activity (~15 Ma; see above). As already discussed for the eastern Sardinia margin, the major rifting phase in the Tyrrhenian basin occurred in Tortonian-middle Pliocene time (e.g., Trincardi and Zitellini, 1987) and became progressively younger toward the northeast (Bartole, 1995).

Figure 16. Lithospheric scale cross-section from Corsica to the Adriatic (modified after Carminati *et al.*, 2004) and present-day vertical motions.



In this model, the Apennines prism (light green) developed along the retrobelt of the Alpine orogen (the blue double-wedge in the center of the section), where there was oceanic or thinned continental lithosphere able to subduct to the east of the pre-existing orogen. The alpine orogen was boudinated and stretched by the backarc rift. Above the cross section, are reported the vertical movements, their average rate (although lateral variations occur), and the interpreted origin. Note that several geodynamic mechanisms appear to coexist and control the uplift or subsidence along the different segments of the subduction-backarc system.

In the central sectors of the Tyrrhenian Basin, basins bounded by listric normal faults, cutting both Messinian-Quaternary sedimentary sequences and the underlying crust, are filled by Pliocene-Quaternary deposits and lava flows. These faults flatten close to the Moho and locally serpentinized mantle, tectonically overlain by stretched continental crust, crops out (e.g., Kastens, Mascle *et al.* 1987). The Tyrrhenian Sea is floored by igneous rocks in two main areas, the Vavilov and Marsili basins (e.g., Kastens, Mascle *et al.* 1987; Bonatti *et al.* 1990; Beccaluva *et al.*, 1990; Sborshchikov and Al-Mukhamedov, 1992; Trua *et al.*, 2002, 2004, 2007; Peccerillo, 2005; Fig. 11e-f-g-h). In the Vavilov basin ~7.5-3.5 Ma-od volcanic rocks have been dredged and cored, and two large volcanic structures have been identified: the Magnaghi seamount (~3.0-2.7 Ma) and the Vavilov seamount (~0.7-0.1 Ma; Trua *et al.*, 2004, 2007). In the Marsili basin (~1.9-1.6 Ma; Fig. 11g) only a giant volcano has been identified, the Marsili seamount (<0.7 Ma; Fig. 11h). The composition of the Vavilov and Marsili seamounts range from T- (Transitional) to E-type (Enriched) MORB to calcalkaline basaltic and high-K andesitic, being N-MORB lithologies nearly absent (Fig. 10b). Sodic alkaline rocks (mostly hawaiites) have been dredged only along the Magnaghi seamount flanks, on the Quirra seamount (SW Tyrrhenian Sea) and rarely along the Vavilov seamount (Beccaluva *et al.*, 1982; Robin *et al.*, 1987; Peccerillo, 2005).

North of NE Sicily, seven major volcanic islands (Aeolian Islands: Filicudi, Alicudi, Panarea, Lipari, Vulcano, Salina and Stromboli: Fig. 11h) plus a large number of volcanic seamounts are active from ~1.3 Ma (e.g., De Astis *et al.*, 2000; Calanchi *et al.*, 2002; Gioncada *et al.*, 2003; Peccerillo, 2005; Francalanci *et al.*, 2004, 2007). Two of the seven volcanoes (Stromboli and Vulcano) are currently in activity. The Aeolian Islands are located on a 15-20 km-thick continental crust, above an active and seismogenic NW steeply dipping (70-85°) Benioff-Wadati plane (e.g., Chiarabba *et al.*, 2008 and references therein). The igneous activity of the Aeolian Islands started with calcalcaline products (Sisifo seamount), continued with shoshonitic activity (Eolo and Enarete seamounts) and then ultrapotassic magmatism developed on the islands, together with the previous two magmatic series (Francalanci *et al.*, 2007 and references therein; Fig. 11b). A general increase of potassium content of the magma is recorded in several islands (e.g., Vulcano and Stromboli) but this is not a rule for the entire district (e.g., Peccerillo, 2005; Francalanci *et al.*, 2007).

Ustica is a small volcanic island located on a ~50 km-thick lithosphere along the southern margin of the Tyrrhenian Sea for a total height of ~2220 m (with less than 10% sub-aerial; Fig. 11h). It is made up of Quaternary volcanic rocks (0.75-0.13 Ma; De Vita *et al.*, 1998) with mildly Na-alkaline affinity. The rocks range in composition from alkali basalts to rarer trachytes (Cinque *et al.*, 1988; Bellia *et al.*, 2000; Trua *et al.*, 2003; Peccerillo, 2005; Fig. 10a). Interestingly, after a strong earthquake occurred in 2002 few km off-shore the NW coast of Sicily (about 40 km SE Ustica), the communications through the fibre-optic cable between Palermo and Roma were interrupted. After the recovery of the cable, it was observed that a side was completely burnt and that about 3 km of cable were locked. Favali et al (2006) interpreted this feature as the consequence of a volcanic eruption of a new active and still unknown volcano.

Sicily Channel

In the Sicily Channel the Apennines-Maghrebides belt is cross-cut by a NW-trending rift system. Three main elements mark the accretionary prism of Sicily and its offshore continuation (Corti *et al.*, 2006): (a) a Pliocene–Pleistocene northwest-dipping foreland monocline generating the overlying foredeep (Mariotti and Doglioni, 2000); (b) a roughly ENE-WSW to E-W-trending thin-skinned imbricate wedge, progressively emplaced from the Early-Middle Miocene to Present (e.g., Roure *et al.*, 1990a; Catalano *et al.*, 1996); (c) northwest-trending normal faults and related grabens or half grabens, associated with a Pliocene-Recent rifting phase (e.g., Reuther and Eisbacher, 1985; Torelli *et al.*, 1995) that led to the development of the Sicily Channel. Toward the northwest, the Sicily Channel rift seems to be connected with the Campidano graben in southwest Sardinia (Corti *et al.*, 2006). The rift affects also the Pelagian shelf and onshore Tunisia and possibly continues to the southeast into the Sirte basin in Libya (Corti *et al.*, 2006). South of Sicily the rifting process produced the Pantelleria, Malta, and Linosa grabens.

The islands of Pantelleria and Linosa are the main volcanoes of the Sicily Channel district (Fig. 11h). These are located in a transtensive area of post-Miocene NW-trending horsts and grabens between Sicily and Tunisia. The igneous activity of Pantelleria (~118-4 ka) is younger than that of Linosa (1.06-0.53 Ma; Rossi *et al.*, 1996). Both the islands are characterized by rocks with sodic alkaline compositions, with type-locality of pantellerites (peralkaline rhyolite) at Pantelleria (Esperança and Crisci, 1995; Civetta *et al.*, 1998; Avanzinelli *et al.*, 2004). The volcanic products of the island of Linosa are nearly indistinguishable from the Pantelleria rocks, but mafic to intermediate compositions are here more abundant than evolved lithotypes (Di Bella *et al.*, 2008; Fig. 10a). The Sicily Channel district is also characterized by the presence of several volcanic seamounts (e.g., Graham and Nameless Banks, Tetide, Anfitrite, Galatea and Cimotoe seamounts, plus other minor volcanic centers; e.g., Beccaluva *et al.*, 1981; Calanchi *et al.*, 1989; Rotolo *et al.*, 2006). Rock types are, in order of abundance, hawaiites, alkali basalts, basanites, mugearites and tholeiitic basalts (Fig. 10a).

What we imagine: the geodynamic evolution

The pieces of information presented in the above sections are integrated in a geodynamic scenario dominated by the convergence of Africa against Europe. In the Italian area, three subduction zones active during the Cenozoic consumed both oceanic and continental lithosphere (Doglioni and Carminati, 2002): the Alpine (with the European Plate subducting below the Adriatic micro-plate), the Apenninic (with the nearly completely consumed Ionian/Mesogean oceanic lithosphere and the Adriatic micro-plate subducting westward beneath the European Plate and characterised by a strong “eastward” radial retreat of the subduction hinge), and the Dinarides (with the Adriatic micro-plate subducting north-eastward under the European Plate). The third subduction, that is marginal in the Italian region, is no longer treated in this review. The following geodynamic reconstruction is shown in the movie associated with this paper, downloadable from the Journal of Virtual Explorer repository system.

The movie: some information and caveats

Although a full and detailed description of the evolution of the entire western and central Mediterranean region is beyond the scope of this work, the map-view movie attached to this paper includes also features external to the Italian area (e.g., Pyrenees, Dinarides, Betics, Rif, Maghrebides, Carpathians, Pannonian Basin), not treated in this work. This was done in order to place the geodynamic evolution of the Italian region in a more regional context. For more details on the evolution of the Mediterranean region and for alternative views, the interested reader is referred to previous reviews of the geology and geodynamics of the central and western Mediterranean area (e.g., Faccenna *et al.*, 1997. Carminati *et al.*, 1998; Gueguen *et al.*, 1998; Wortel and Spakman, 2000;

Tari, 2002; Csontos and Voros, 2004; Carminati and Doglioni, 2005; Rosenbaum and Lister, 2004; Mauffret, 2007; Chalouan *et al.*, 2008).

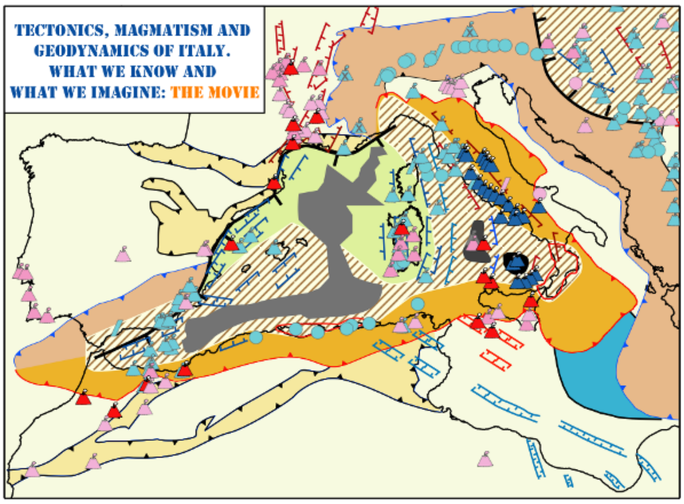
The movie that is here presented is intended to be as rigorous as possible. However, the scale of the representation and the complexity of the evolution of the region are such that some features had to be represented in a schematic way. For example, the normal faults associated with the opening of the back-arc basins are not to be taken as representative of real faults but indicate the location where normal faulting occurred. Moreover, we have to admit the following uncertainties and under-constraints in the paleogeographic reconstructions. A general (and easily understandable) rule is that the uncertainties are larger for older time frames. In the 50-40 Ma frames of the movie, the geometry of the Liguro-Piedmont-Penninic ocean is largely speculative.

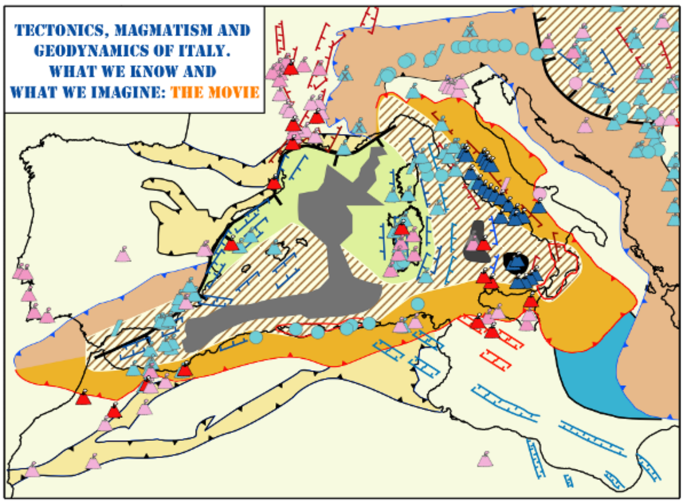
Another poorly constrained point is the contact between the Betics and Rif mountain belts, that, in the movie is idealized with a straight line. As concerns the age of the igneous activity, most of the geochronological data are incomplete and scattered (no systematic cover exists). Most of the isotopic ages are based on old (produced during ‘70s-‘80s) K-Ar ages and only very few are based on detailed $^{40}Ar/^{39}Ar$ ages. Moreover, as in the Tyrrhenian Sea submerged igneous rocks, only the most recent activity has been dredged/cored. This means that the ages should be considered only as the youngest limit. No one knows exactly when igneous activity started in these and other igneous districts.

In order to clarify the evolution of the area, a section-view movie showing schematically the evolution of the area is also provided. A PDF file with the animation can be downloaded from the JVE website - Download The Movie PDF.

Many features of the movie (e.g., fronts of Alps and Dinarides) are a function of plate kinematics reconstructions. Different kinematic reconstructions would imply different locations of these fronts. In the movie, Europe is kept fixed and behaves as a single plate, while Africa and Adria are moved coherently. This choice is largely under-constrained and implies that: 1) no deformation occurs within the Ionian lithosphere; 2) the rifting/drifting that drove the development of the Ionian lithosphere was completely finished at the beginning of the movie (for alternative paleogeographic reconstructions, the readers are referred to Muttoni *et al.*, 2001; Bosellini, 2002; Conti *et al.*, 2005); and 3) well know transcurrent features, such as the Mattinata and Tremiti faults (e.g., Tondi *et al.*, 2005), are neglected. Also the shape and lateral continuity of the Ionian oceanic lithosphere is highly speculative.

Figure Animation. The Movie - plan view and section view.





Quicktime .mov files. Plan view (top) and section view (bottom).

The geometry of the link between Dinarides and Alps is perhaps the most under-constrained feature in this movie. The ancient (Paleogene-beginning of Neogene) intersection between the two mountain belts is now buried under the sediments of, or scattered within discontinuous outcrops in the Pannonian Basin. This makes it difficult to establish whether Alps and Dinarides were linked during Paleogene times. The answer is related to the choice of the paleogeography of the Mesozoic oceans that developed between Africa and Eurasia. If the oceans subducted in the Alps and Dinarides subduction zone were connected (e.g., Schmid *et al.*, 2008), then the two belts were likely linked since the very beginning of the collisional stage. If the oceans were disconnected (e.g., Stampfli *et al.*, 2002; Stampfli and Hochard, 2009), then no ancient connection between the two belts is necessary.

Also the location of the retreating front of the Apennines-Maghrebides and the distance between the active front and the hinge of the subducting slab are largely under-constrained. At present, geophysical and geological data suggest that this distance can vary considerably from salients to recesses of the belt. In addition, the present-day front of the various belts is highly segmented. The front was likely segmented also in the past, although the fronts are shown as linear features in the movie.

Alps

After a long period of divergence, associated with Mesozoic rifting along the Hercynian paleo-suture (Santantonio and Carminati, 2010) and drifting of the Liguro-Piedmont-Penninic Ocean (a branch of the larger Tethyan Ocean; e.g., Stampfli *et al.*, 2002), the convergence process between the African and Eurasian plates started at the end of the Early Cretaceous or at the beginning of the Late Cretaceous, according to different reconstructions (e.g., Smith, 1971; Dewey *et al.*, 1973; Dewey *et al.*, 1989; Polino *et al.*, 1990; Rosenberg *et al.*, 2002). Along the present Alpine area, during the Cretaceous-Paleocene, the consumption of the Tethyan ocean(s) (see Polino *et al.*, 1990, Schmid *et al.*, 1996, Scotese, 1991, Stampfli *et al.*, 2002, for a discussion on Cretaceous palaeogeography) took place. Subduction is also testified by HP-LT metamorphism recorded in ophiolites and former passive margin rocks in the Alps (Spalla *et al.*, 1996), in the northern Apennines (Rossetti *et al.*, 1999) and in Calabria (Liberi *et al.*, 2006). Eclogitic and blueschist facies metamorphism in the Western Alps was recognised as an evidence of a fossil subduction zone (Ernst, 1971), soon after the development of the plate tectonics theory and leading to the reinterpretation of the Alpine-Himalayan orogenic belt as the result of subduction and collision processes (Laubscher, 1969, 1970; Dewey and Bird, 1970). The eclogite facies metamorphism in the Sesia-Lanzo (Austroalpine) and Monte Rosa (Penninic) basement slices was soon recognized as a proof that large, coherent slices of continental crust were subducted (Dal Piaz *et al.*, 1972; Marchant and Stampfli, 1997).

Interestingly, no undisputable syn-subduction (i.e., pre-collision) igneous activity is recorded along the Alpine chain. This feature (absence of syn-subduction igneous activity) is similar to what happened in the neighbouring and coeval Pyrenean Chain. The only syn-subduction volcanic activity, located in the Veneto region (NE Italy; Fig. 11a), shows “anorogenic” geochemical characteristics that are not compatible with a subduction-related origin and are rather inconsistent with the regional geodynamic setting – the foreland of the Southern Alps. Hereafter, the term anorogenic is used following the rationale proposed by Wilson and Bianchini (1999), Lustrino and Carminati (2007) and Lustrino and Wilson (2007). The term “orogenic” is similarly based on the rationale proposed by Wilson and Bianchini (1999) and Lustrino *et al.* (2011). What we want to stress is that geochemical terms cannot automatically be exported as geological concepts. There is no ultimate and absolutely valid geochemical or mineralogical tool to be used to define a given igneous rock as orogenic (or “subduction-related”) or anorogenic (or “intra-plate-like”). Typically, but with several exceptions, the anorogenic and orogenic igneous rocks can be distinguished considering the following geochemical characteristics: a) anorogenic rocks show lower SiO_2 contents (typically <50wt.%) than orogenic rocks (typically >45 wt%); b) anorogenic rocks typically have K_2O/Na_2O ratios < 1.5 (commonly <1), whereas orogenic rocks have K_2O/Na_2O ratios from 1 to >10; c) anorogenic rocks have typically TiO_2 >1.2 wt. %, while orogenic rocks have TiO_2 <1.5 wt.%; d) anorogenic rocks have relatively lower LILE (Large Ion Lithophile Elements) content compared to orogenic rocks; e) anorogenic rocks have lower LILE/HFSE (High Field Strength Elements) ratios and higher HFSE contents than orogenic rocks; f) anorogenic and orogenic rocks have typically $^{87}Sr/^{86}Sr$ isotopic ratios lower and higher than BSE (Bulk Silicate Earth estimate = 0.70445), respectively; g) anorogenic and orogenic rocks have typically $^{143}Nd/^{144}Nd$ isotopic ratios higher and lower than ChUR (Chondritic Uniform Reservoir estimate = 0.51264); h) anorogenic rocks have slightly higher $^{208}Pb/^{204}Pb$ for a given $^{206}Pb/^{204}Pb$ ratio compared to orogenic rocks; i) anorogenic rocks have $^{207}Pb/^{204}Pb$ typically <15.66 (down to ~15.49) compared to values of >15.65 (up to 15.71) of orogenic rocks; j) anorogenic rocks have typically $\sigma^{18}O$ values around 5.2-5.5‰, lower than typical values of orogenic rocks (up to 8-10). For a more complete treatment of these arguments the readers are referred to Lustrino *et al.* (2011).

It has been claimed in the literature (e.g., Polino *et al.*, 1990) that indirect evidence for subduction-related volcanism is preserved in the volcanic clast-rich turbidites of the Taveyanne Sandstone Formation (Savoie, France; Fig. 11b-c), originally attributed to Eocene (43 ± 4 Ma; Vaugnat, 1983). However, later geochronological ^{40}Ar/^{39}Ar datings on volcanic amphiboles and biostratigraphic dating on calcareous nannofossils associated with these rocks yielded ages of ~32-29 Ma (Ruffini *et al.*, 1997 and references therein), unequivocally proving that this volcanism was syn-collisional (i.e., occurred after the end of the subduction process). Given its position and the lateral continuity with the coeval Provence volcanism, the Taveyanne volcanism could also be related to the Apenninic subduction, that will be later discussed.

Until the Eocene, the Alpine chain extended from present day Alps to the Betic Chain in SE Spain (see the movie). The continuity of the belt can be reconstructed restoring to their original position the Corsica, Calabria, Peloritani Mts. and Kabilies terranes, that display clear evidence of Alpine deformation in terms of age and tectonic transport (Michard *et al.*, 2006; Molli, 2008; Vignaroli *et al.*, 2009; Heymes *et al.*, 2010, and references therein). The Europe-Adria collision took place diachronously during Eocene-Oligocene? time (Stampfli et al., 1998; Handy et al., 2010), after the complete subduction of the Ligurian-Piedmont-Penninic Ocean beneath the Adriatic continental lithosphere (Schmid *et al.*, 1996). Strain rate decreased after the onset of collision due to decreasing convergence rates (Schmid *et al.*, 1996). The collision brought to a phase of regional Barrowian metamorphism, characterized by high grade in the axial parts of the belt (e.g., Lepontine Dome and Tauern Window) and by decreasing grade moving toward the outer parts of the fold-and-thrust belt. As discussed, exhumation of such metamorphic rocks occurred with different patterns in Eastern (stationary exhumation) and central-western (migrating exhumation) Alps. This different behaviour has been related to north directed, lower crustal wedging in the middle and upper crust of the Central and Western Alps. The absence of such lower crustal wedge in the Eastern Alps is possibly associated to the smaller shortening and consequent smaller exhumation in the same area. In other words, the deep structure and rheology of the orogen, associated with the total shortening, and not lateral changes in its erosional efficiency, provided the main control on the lateral growth of the Alps (Rosenberg and Berger, 2009).

The continued convergence between European and Adriatic plates during post-collisional time led to accumulation of shortening and continuous enlargement of the double vergence belt (Fig. 17), with the involvement of Mesozoic and Tertiary sediments belonging to previous European and Adriatic passive margins and foredeep and foreland basin (Po Plain and Molasse basins) sediments.

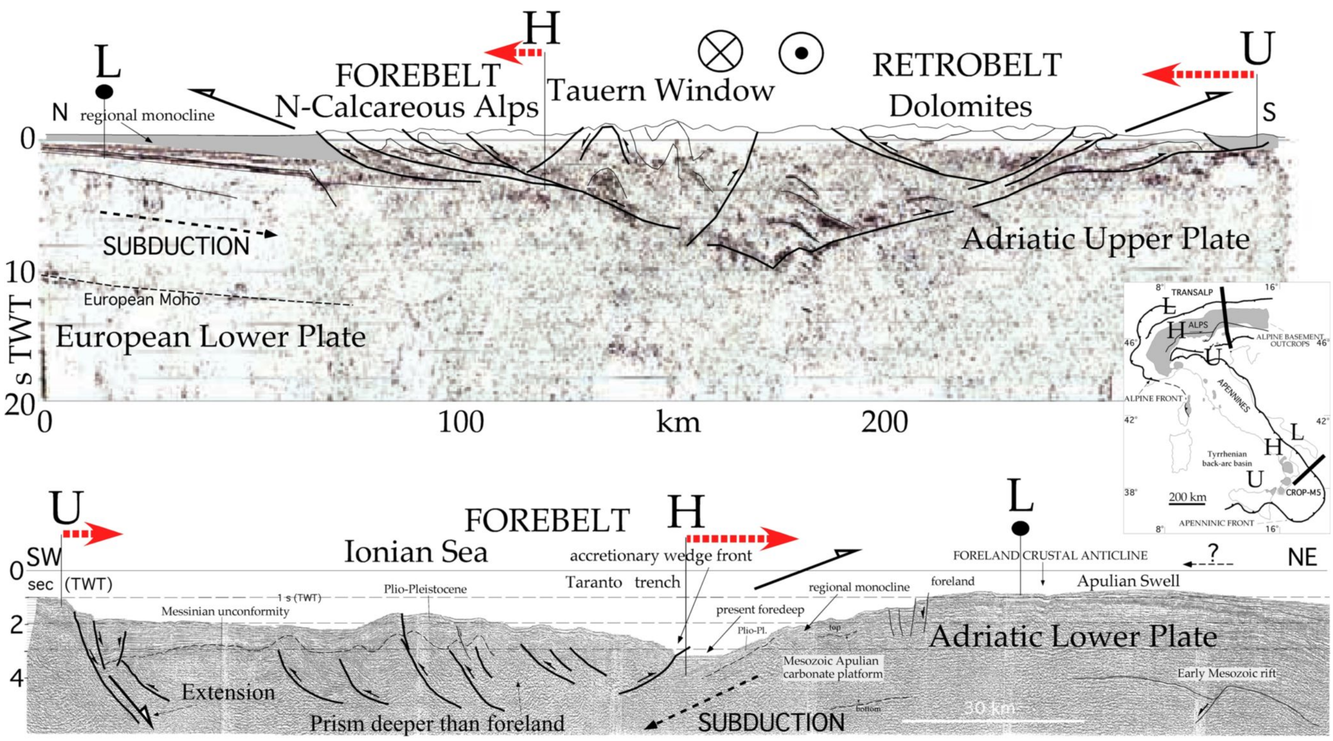
In the Early Oligocene (~33-29 Ma) the Alps were characterized by a sharp increase of igneous activity with emplacement of abundant calcalkaline plutonic and subvolcanic rocks (Fig. 11b). These rocks are generated by both mantle (only Adamello gabbros, essentially) and hybrid mantle + crustal sources (the bulk of the plutons; e.g., von Blanckenburg and Davies, 1995), as evidenced by their variable ^{87}Sr/^{86}Sr initial isotopic ratios (~0.704-0.716) and ^{143}Nd/^{144}Nd initial values (~0.5128-0.5121; Fig. 18). The Periadriatic-Insubric plutons show Δ7/4 and Δ8/4 values (Fig. 19a) ranging from ~+4 to ~+80 and from ~+30 to ~+87, respectively, a feature that is commonly related to recycling of ancient sediments (Hawkesworth *et al.*, 1986), delaminated lithospheric mantle (e.g., Mahoney *et al.*, 1992) or lower continental crust (e.g., Escrig *et al.*, 2004).

The Δ7/4 and Δ8/4 values represent the vertical distance of ^{207}Pb/^{204}Pb and ^{208}Pb/^{204}Pb, respectively, compared to a given ^{206}Pb/^{204}Pb value lying on the Northern Hemisphere Reference Line (NHRL) in^{207}Pb/^{204}Pb vs. ^{206}Pb/^{204}Pb and ^{208}Pb/^{204}Pb vs. ^{206}Pb/^{204}Pb isotopic spaces, respectively. The NHRL is considered to represent the Pb isotopic composition of oceanic basalts of the northern hemisphere of the Earth (Hart, 1984).

What is the origin of the Insubric plutons and why no subduction-related igneous activity is recorded along the Alpine Chain? The slab break-off hypothesis - that associates the magmatism with the rupture of the slab subducting under the Alps, a process that eventually led to the upwelling af mantle and to its partial melting - is the most accepted petrogenetic model to explain the igneous activity along the Periadriatic Line. Von Blanckenburg and Davies (1995) proposed this model to explain the linear distribution of the Periadriatic magmatism, the short and nearly coeval timing of emplacement of nearly all plutons, and the mantle-related source of the magmas.

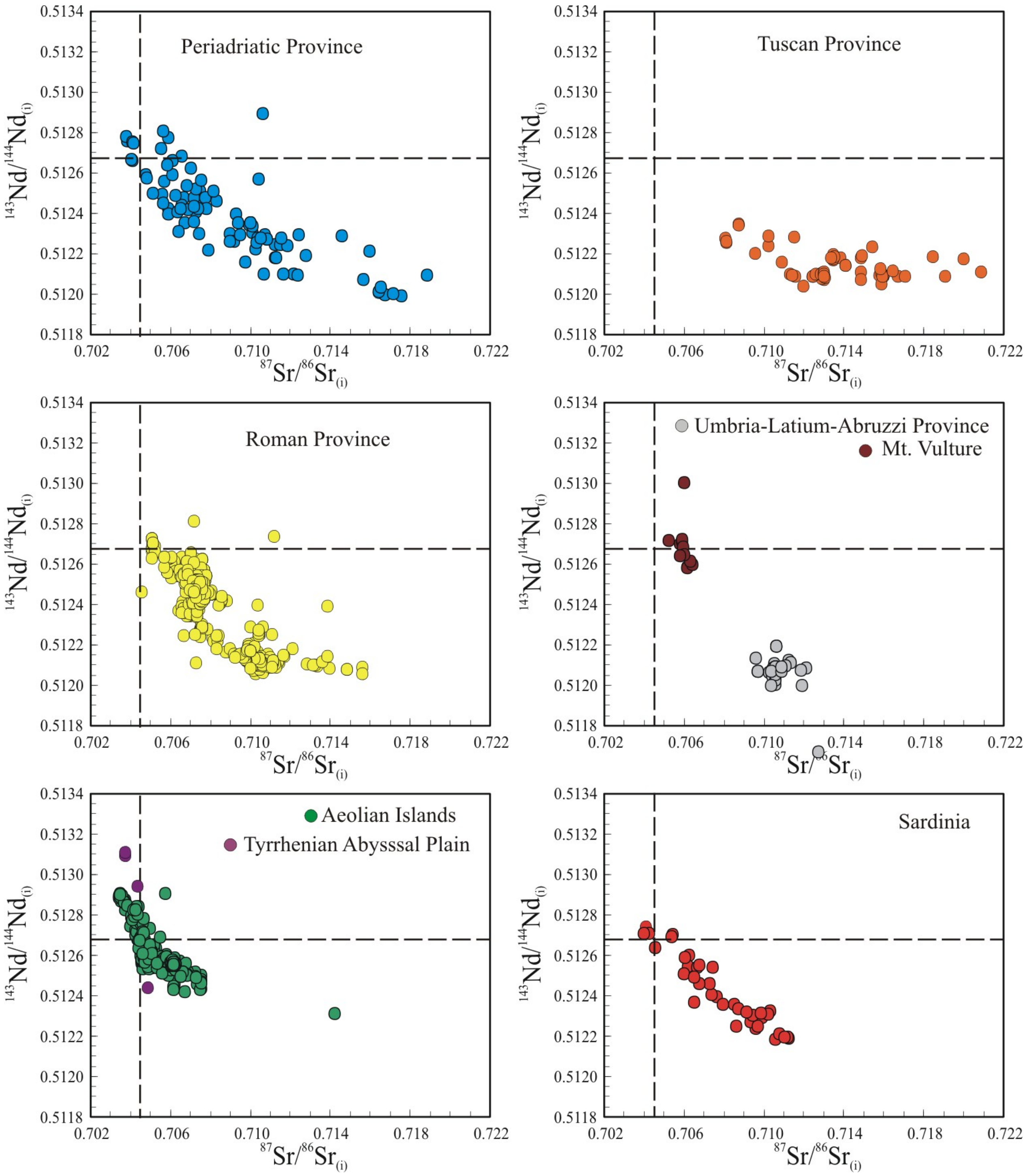
However, it has been argued that the linear distribution in map view does not necessarily reflect the shape of the source region, but only that of a tectonically-controlled ascent pathway (Rosenberg, 2004). The nearly horizontal rupture in the down-going slab would occur when relatively light continental lithosphere enters the trench, after continental collision. This would produce opposing buoyancy forces at depths creating strong extensional forces within the slab, ultimately tearing off and detaching the dense oceanic slab (von Blanckenburg and Davies, 1995). Based on the results of seismic tomography, Lippitsch *et al.*(2003) observed a spatial gap in the slab presently subducted below the Alpine orogen, interpreting this as the evidence for the slab break-off occurred during Late Eocene-Early Oligocene times. However, this interpretation is based on a 1D P-waves velocity model and can be questioned on the basis that continental subduction has been documented in the Alps down to 150-200 km (Panza *et al.*, 1992). Moreover, the occurrence of the detachment in the tomographic models is a function of the adopted velocity model. The dehydration of the European downgoing continental segment of the slab could alternatively account for 1) the melting of the upper plate lithospheric mantle, 2) the crustal contamination of magmas, and 3) the slightly seismically slower lithospheric mantle in the upper plate, which may mimic a slab detachment (e.g., Doglioni *et al.*, 2009).

Figure 17. Seismic cross sections through Alps and Apennines (modified after Doglioni *et al.*, 2007).

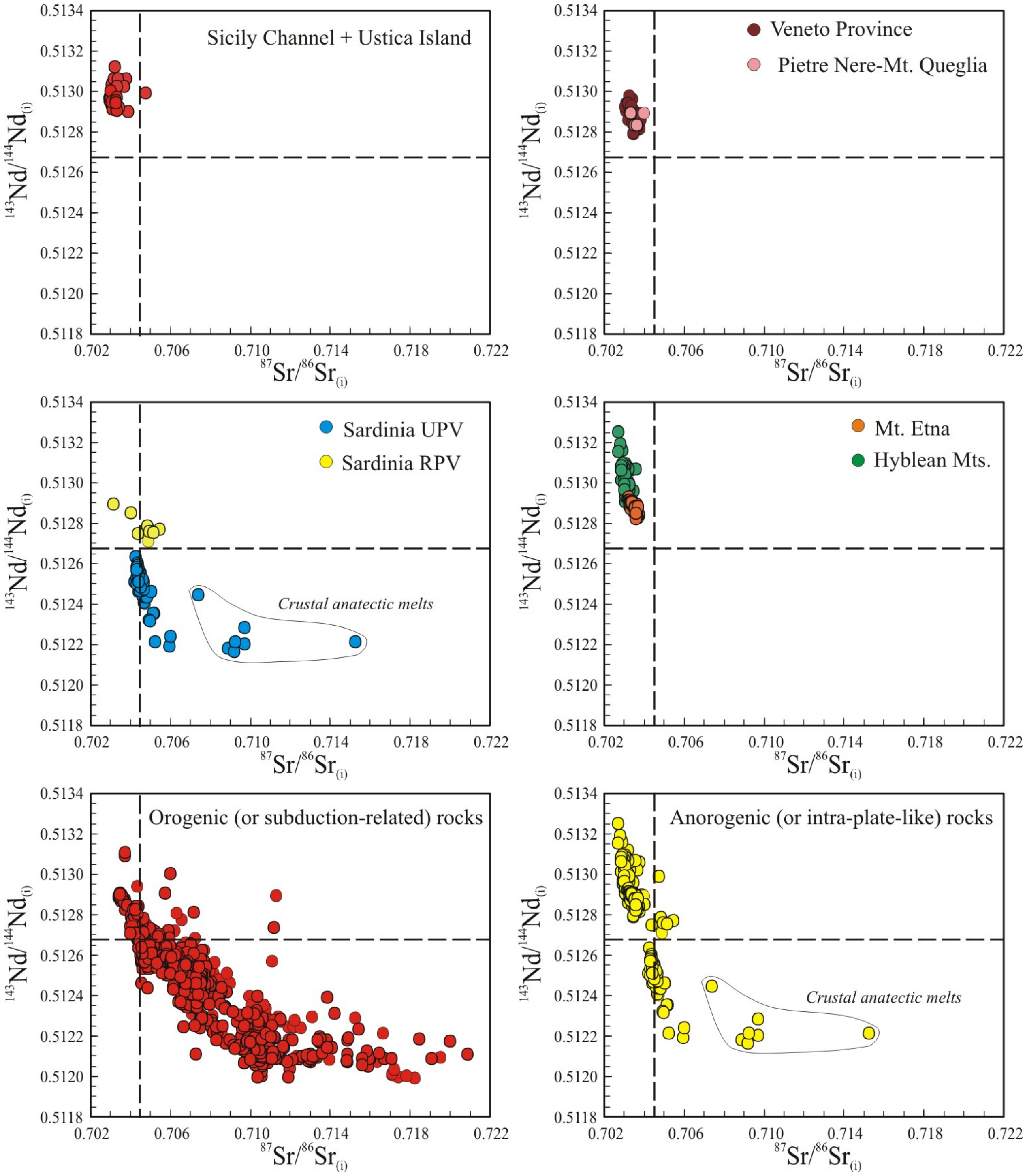


Note the Alpine double vergence (upper section), whereas the Apennines (lower section) have a single vergence. The Apennines prism is deeper than the foreland and characterized by extension at its rear (to the left). L, lower plate, H, subduction hinge, U, upper plate. Assuming fixed the lower plate (Europe for the Alps, and the Adriatic plate for the Apennines), the subduction hinge the subduction hinge is advancing towards the upper plate in the Alps and retreating in the Apennines.

Figure 18. Sr-Nd isotopic ratios of the Italian Cenozoic igneous rocks

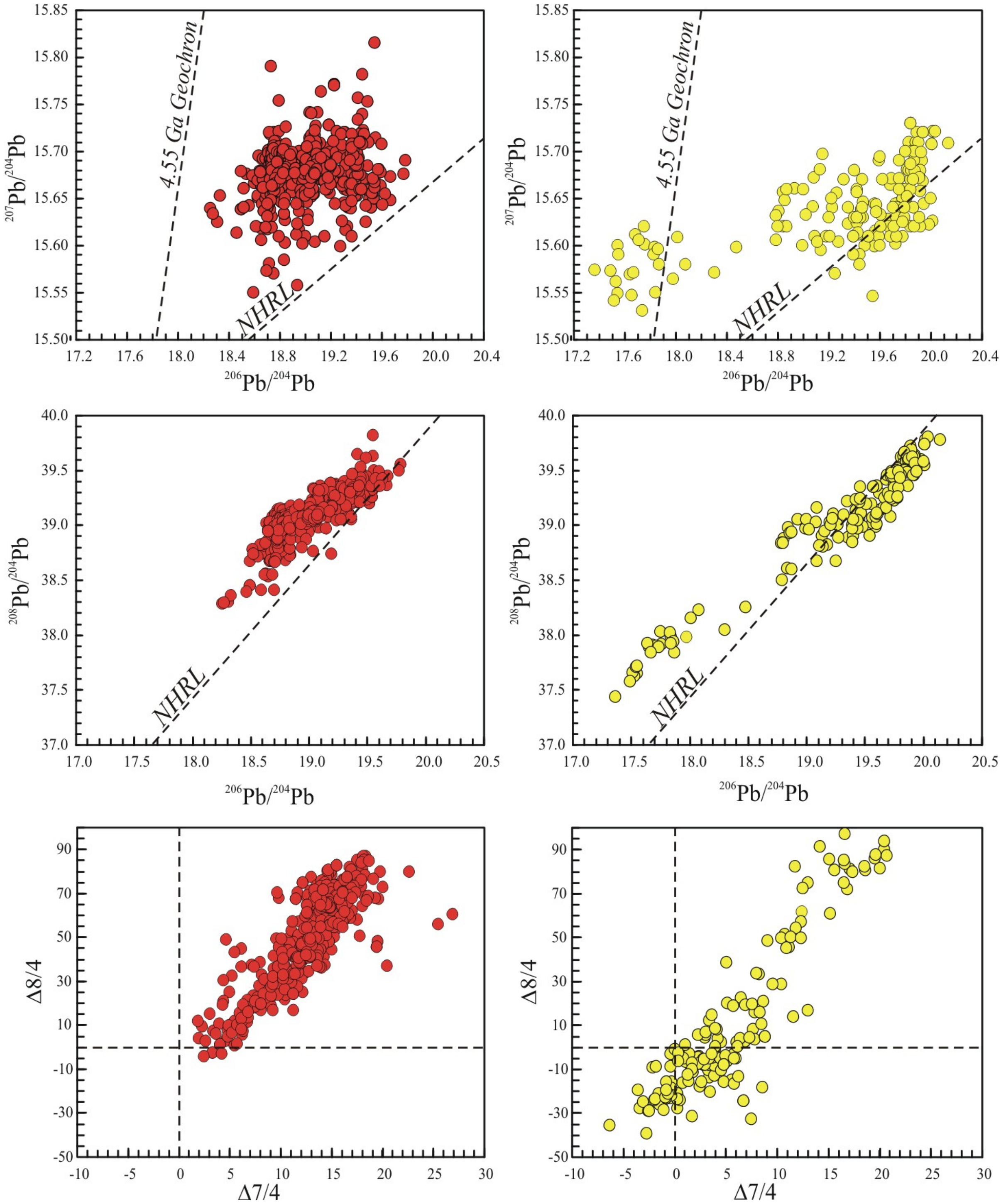


Top (a) = $^{143}Nd/^{144}Nd$ vs. $^{87}Sr/^{86}Sr$ initial isotopic ratios for the Cenozoic subduction-related (or "orogenic") igneous rocks of Italy. BSE = Bulk Silicate Earth estimate ($^{87}Sr/^{86}Sr$ = 0.70445). ChUR = Chondritic Uniform Reservoir estimate ($^{143}Nd/^{144}Nd$ = 0.51264).

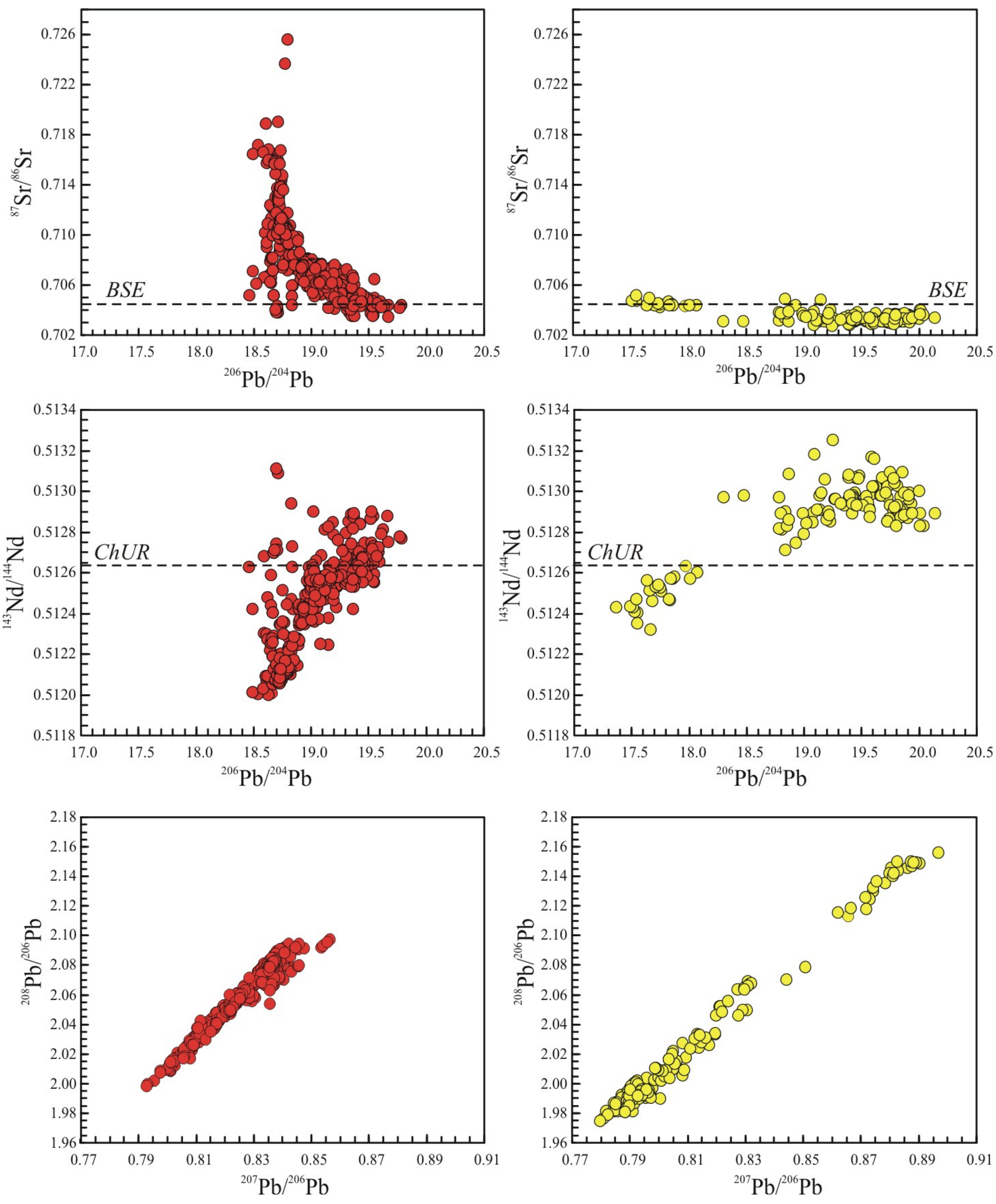


Bottom (b) = $^{143}Nd/^{144}Nd$ vs. $^{87}Sr/^{86}Sr$ initial isotopic ratios for the Cenozoic intra-plate-like (or "anorogenic") igneous rocks of Italy. The samples enveloped in the thin continuous line are crustal anatectic rhyolites and trachytes from Mt. Arci (Sardinia) considered pure crustal (not mantle) partial melts. References for subduction-related rocks in Lustrino *et al.* (2011). References for intra-plate-like rocks in Lustrino and Wilson (2007) plus updates downloadable from the GEOROC web site (http://www2.mpch-mainz.mpg.de/~geo/Databases/GEOROC/Expert_Datasets.htm). The full list of analyses and references can be requested to the second author.

Figure 19. Pb-Sr-Nd isotopic ratios of the Italian Cenozoic igneous rocks

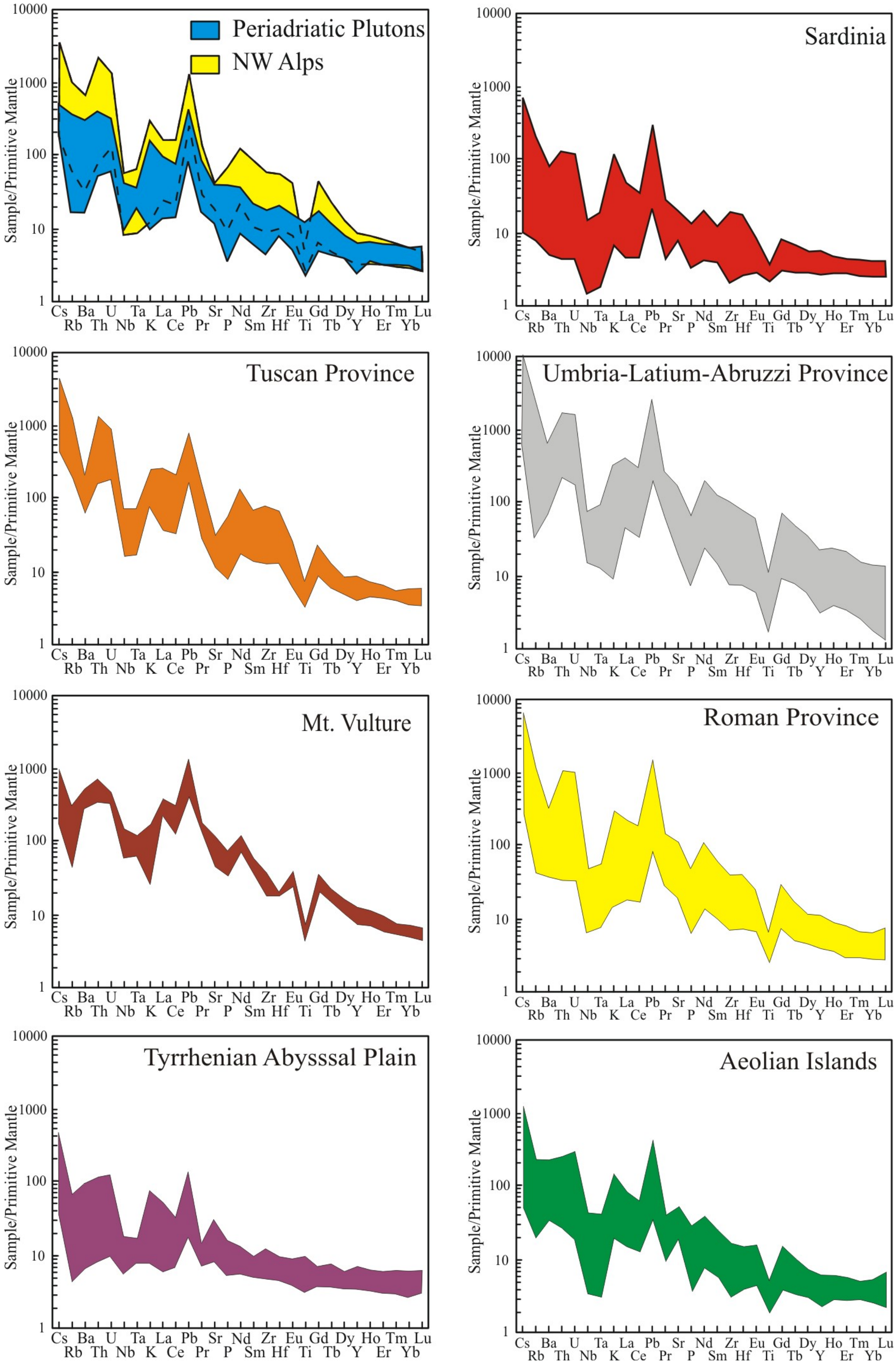


Top (a) = $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$, $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ and Δ8/4 vs. Δ7/4 diagrams for subduction-related (red circles) and intra-plate-like (yellow circles) Cenozoic igneous rocks of Italy. Δ8/4 and Δ7/4 parameters reflect the vertical shift of $^{208}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ from the NHRL (Northern Hemisphere Reference Line; Hart, 1984) for a given $^{206}Pb/^{204}Pb$, respectively. The NHRL is plotted for reference in the first four diagrams. Also shown the 4.55 Ga Geochron.

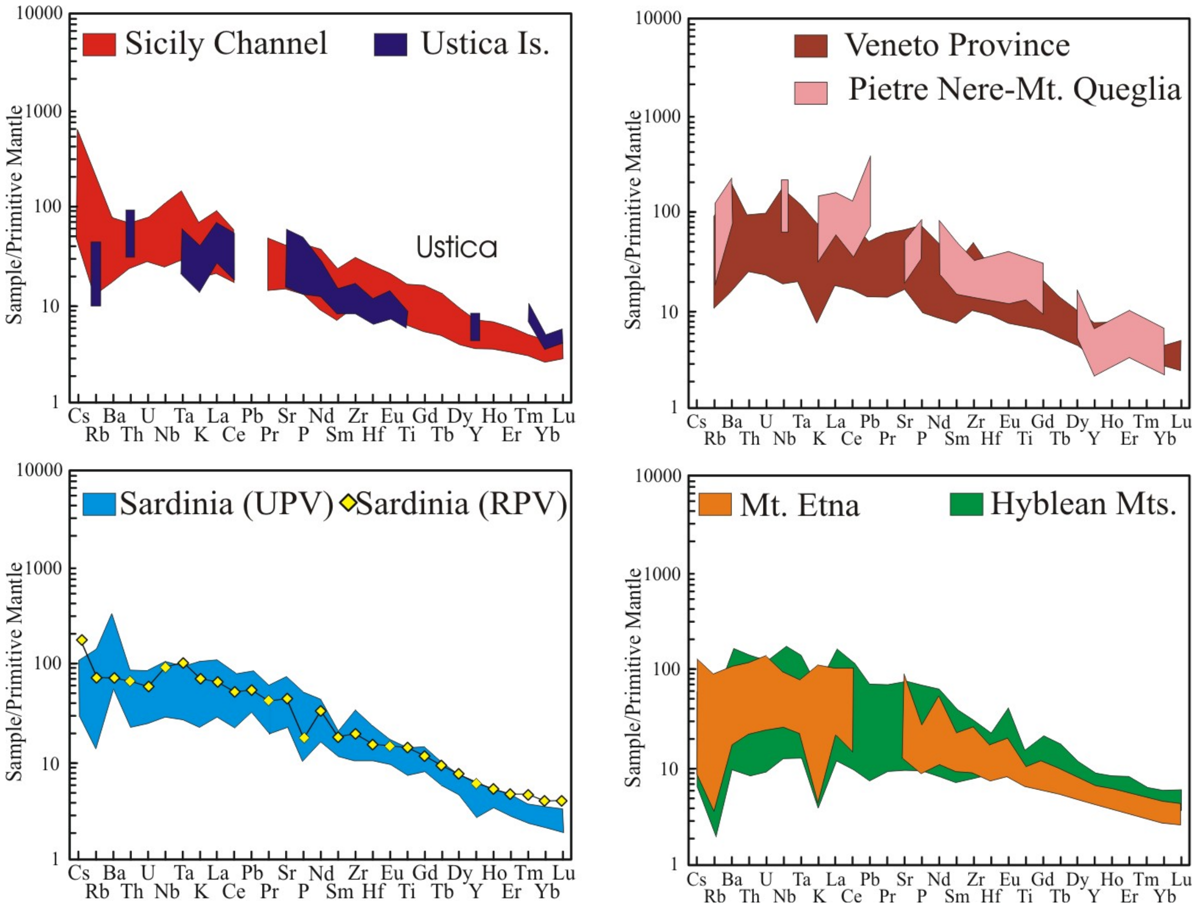


Bottom (b) = $^{87}Sr/^{86}Sr$ vs. $^{206}Pb/^{204}Pb$, $^{143}Nd/^{144}Nd$ vs. $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{206}Pb$ vs. $^{207}Pb/^{206}Pb$ diagrams for subduction-related (red circles) and intra-plate-like (yellow circles) Cenozoic igneous rocks of Italy. References for subduction-related rocks in Lustrino *et al.* (2011). References for intra-plate-like rocks in Lustrino and Wilson (2007) plus updates downloadable from the GEOROC web site (http://www2.mpch-mainz.mpg.de/~geo/Databases/GEOROC/Expert_Datasets.htm). The full list of analyses and references can be requested to the second author.

Figure 20. Primitive mantle-normalized multielemental diagram of the most mafic Italian Cenozoic igneous rocks



Top (a) = subduction-related (or “orogenic”) rocks. References in Lustrino *et al.* (2011).



Bottom (b) = intra-plate-like (or “anorogenic) rocks. References in Lustrino and Wilson (2007) plus updates downloadable from the GEOROC web site (http://www2.mpch-mainz.mpg.de/~geo/Databases/GEOROC/Expert_Datasets.htm). The full list of analyses and references can be requested to the second author.

Less clear is the absence of subduction related igneous activity during the subduction of Tethyan oceanic lithosphere. Von Blanckenburg and Davies (1995) explained this feature as consequence of the slow pre-Oligocene convergence rates between Adria and Europe (<1 cm/yr).

The rare dykes intruding the Periadriatic plutons show commonly exotic chemical and mineralogical composition, with K content increasing from the SE Alps (mostly tholeiitic to calcalkaline compositions) to the Central Alps (mostly high-K calcalkaline) and NW Alps (mostly shoshonitic to ultrapotassic; Beccaluva *et al.*, 1983). The NW Alps ultrapotassic dykes show a peculiar composition being mostly mafic (e.g., MgO ~6-16 wt%; high Cr and Ni content) but showing Sr-Nd isotopic composition typical of continental rocks. The paradox is that their MgO, Cr and Ni content indicate equilibrium with peridotitic assemblages, but their isotopic signal ($^{87}Sr/^{86}Sr$ ~0.706-0.719; $^{143}Nd/^{144}Nd$ ~0.5120-0.5124; Fig. 19a) reflects a crustal source. All the NW Alpine dykes (high-K calcalkaline, shoshonitic and lamprophyric/lamproitic) show a typical enrichment in Rb, Th, U, Pb and LREE and depletion in Nb, Ta, Hf, Zr and Ti in primitive mantle-normalized diagrams (Fig. 20a). These geochemical characteristics allowed to infer an anomalously enriched mantle source for these magmas related to sediment input into the mantle (e.g., Venturelli *et al.*, 1984; Peccerillo and Martinotti, 2006; Owen, 2008; Conticelli *et al.*, 2009). Recycling of sediment (and upper crustal lithologies in general) should have happened during the eo-Alpine phase (Early Cretaceous-Middle Eocene) prior to the Europe-Adria collision.

A further complication to the evolution of the Alpine collision has been proposed for the Eastern Alps. In the Tertiary, strike-slip tectonics was abundant in this area, and was ascribed to an eastward lateral extrusion process (e.g., Ratschbacher *et al.*, 1991). As an alternative, it is proposed that strike-slip tectonics may be explained by the interplay between the Alpine, Dinaric subductions and the transtension associated with the Pannonian rift development.

Alpine compressional tectonics recorded in the offshore east of Sardinia (Sartori, 2005) in north-eastern Corsica (Molli, 2008), Liguria (NW Italy; Vignaroli *et al.*, 2008) and in Calabria-Peloritani Mts. (Vignaroli *et al.*, 2009; Heymes *et al.*, 2010) were later superimposed by the western Mediterranean rifting stage associated with the Apennines subduction back-arc. A similar tectonic scenario can be envisaged also for the Eastern Alps. The E-W extensional tectonics in this region may be linked to the back-arc extension associated with the "eastward" retreating Carpathian subduction.

Apennines, Sardinia, Tyrrhenian Sea and Sicily Channel

As shown in Fig. 2, the continental lithosphere of the Adriatic micro-plate is subducting both eastward (beneath the Dinarides) and westward (beneath the Apennines; Carminati and Doglioni, 2005). The Apennines are classically interpreted as the accretionary wedge developed along the subduction hinge of the Adriatic continental and Ionian oceanic plates. The total amount of North-South Africa/Europe convergence at the Tunisia longitude is about 135 km in the last 23 Ma, i.e., about five times smaller than the "eastward" migration of the Apenninic subduction hinge, which moved eastward more than 700 km during the same interval (Gueguen *et al.*, 2008). Therefore the slab retreat cannot be ascribed to the Africa-Europe convergence. It must rather be controlled by the slab-mantle intercation, i.e., the slab pull or, alternatively, to the eastward mantle flow relative to the lithosphere. This "eastward" rollback of the Apenninic subduction produced the back-arc stretching of the lithosphere of the Ligurian-Provençal Basin and the Tyrrhenian Sea, with oceanization in the Vavilov and Marsili subbasins (e.g., Malinverno and Ryan, 1986; Doglioni 1991).

The Apennines subduction zone (with a slab dipping even steeper than 70° along the Calabrian arc; Giardini and Velonà, 1991; Carminati *et al.*, 2002, 2005; Chiarabba *et al.*, 2008) is considered connected, through Sicily and the Sicily Channel, to the North Africa Maghrebides (Rif and Tell Orogens) subduction. Geophysical and igneous petrological considerations provide evidence of the subduction of the Adriatic and Ionian lithospheres underneath the whole Apennines (e.g., Peccerillo, 1985, 2005; Royden *et al.*, 1987; Selvaggi and Amato, 1992; Serri *et al.*, 1993; Selvaggi and Chiarabba, 1995; Piromallo and Morelli, 1997; De Gori *et al.*, 2001; Avanzinelli *et al.*, 2009; Lustrino *et al.*, 2011). Seismic data, moreover, permit to recognize the continental subduction in the north, from the oceanic subduction in the south (e.g., Catalano *et al.*, 2001). The slab has been shown to be continuos and undetached under Calabria (Mele, 1998).

Based on the aforementioned structural and petrologic pieces of evidence, the Paleozoic highly metamorphosed Alpine-age terranes in Calabria and Peloritani Mts. are considered as a part of the Alpine belt that was dismembered during the opening of the Ligurian-Provençal and Tyrrhenian Sea basins from Early Oligocene onwards. The paleo-geographic location of the Calabrian-Peloritani Mts. continental crust is still a matter of debate. It has been proposed to be part of: 1) the European southern paleo-continental margin (Ogniben 1969; Knott, 1987; Dietrich, 1988); 2) the African paleo-margin (Alvarez, 1976; Amodio Morelli *et al.*, 1976); 3) an intervening microplate (Doglioni, 1991; Guerrera *et al.*, 1993, 2005; Cello *et al.*, 1996). The present-day Apennines-related deformation front of the accretionary prism is located offshore in the Ionian basin (Doglioni *et al.*, 1999).

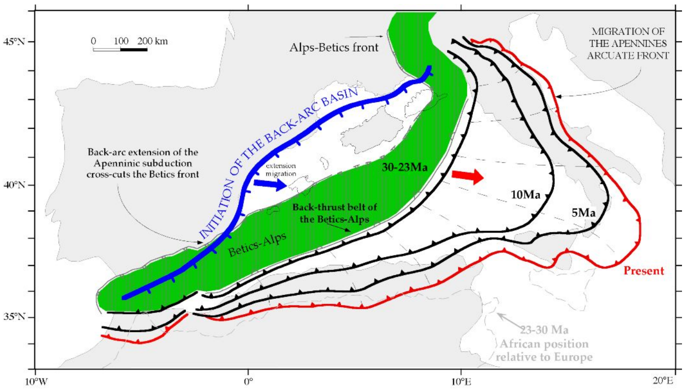
In summary, the Apennines Chain is interpreted as being related to a subduction process involving recycling of continental and oceanic lithosphere and thrusting of different types of lithologies (Meso-Cenozoic carbonatic and deep oceanic basin successions). Evidence for active subduction beneath the Apennines is suggested by local tomographic studies (De Gori *et al.* 2001) and the absence of sub-crustal seismicity within the continental lithosphere of the Adriatic plate subducting under the southern Apennines could be rheologically controlled (Carminati *et al.*, 2002, 2005).

The "westward" subduction and "eastward" retreat of the Ionian oceanic lithosphere and the Adriatic continental lithosphere started probably during Early-Middle Eocene and is largely still active (Devoti *et al.*, 2008). The occurrence of ~38 Ma-old subduction-related volcanic rocks in Sardinia led Lustrino *et al.* (2009) to propose a beginning of subduction between 42 and 49 Ma ago. The age difference between the first magmatic activity and the beginning of the subduction system depends on the relative plate motion, the dipping angle of the subducting slab and the depth at which partial melt occurs in the mantle wedge. The 42-49 Ma age for the beginning of the Apennine subduction system is far older than the 30 Ma age previously proposed (Réhault *et al.*, 1984; Malinverno and Ryan, 1986; Ricci Lucchi, 1986; Patacca and Scandone, 1989; Boccaletti *et al.*, 1990; Gueguen *et al.*, 1998; Carminati *et al.*, 1998). The 30 Ma age was mainly based on previous geochronological estimates available for subduction-related igneous rocks and on the age of syn-tectonic sediments in the western Mediterranean basins. The discrepancy between the age of extensional tectonics and magmatism suggests that the mechanical response to the retreat of the subducting slab is more sluggish than the thermal-petrological response. In addition, such a discrepancy can be also related to the fact that extensional tectonics in the western Mediterranean back-arc basins generally impacted a pre-existing over-thickened lithosphere (Alpine-Betic subduction), apart west of Sardinia, where the rift enucleated in the foreland of the Alps. Thus the first extensional motions, that disrupted the previous topography, were likely associated with the development of irregular continental basins. As a consequence, the sedimentation associated with these early stages may have been discontinuous and the precise dating of such sediments would not be possible.

During Eocene-Oligocene times a critical scenario developed in the proto-western Mediterranean area, with the existence of the S-SE-directed Alpine subduction system (approaching the end of its existence) and the young NW-directed Apennine subduction system (e.g., Carminati *et al.*, 1998; Doglioni *et al.*, 1998; Vignaroli *et al.*, 2008; Lustrino *et al.*, 2009), which possibly developed along the retrobelt of the Alpine orogen (Fig. 21). This means that two subduction systems, with nearly opposite polarity, were present in a relatively narrow area for a short time (Fig. 22).

The southern prolongation of the Western Alps has been incorporated into the internal Apennines, and the Apennines slab retreat induced subsidence and boudinage of large portions of the Alps, partly counteracting their uplift (Fig. 22). From the Marittime Alps southwards, the Alps have been scattered and dismembered into the Apennines-related backarc. However, at least until the Middle Miocene, the Alps continued to be active, in spite of being stretched by the back-arc rift of the ongoing Apennines subduction. This seems testified by the lower Miocene HP/LT rocks of the Tuscan area, an assemblage that can form only where the crust is thickened as in the Alpine system. The coexistence of opposite subduction zones is documented also in other areas, like central America and Molucche.

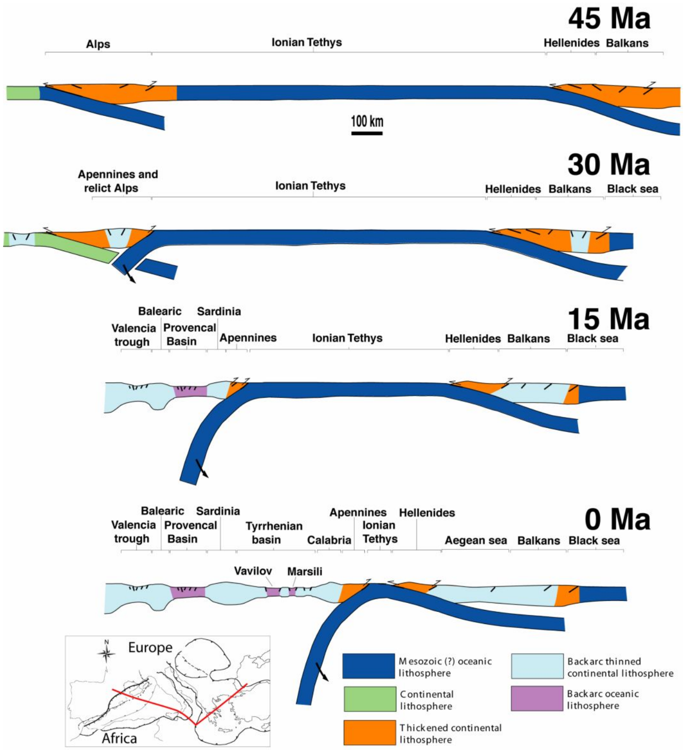
Figure 21. The W-directed Apenninic subduction started in the western Mediterranean in the Eocene(?)-Early Miocene along the retrobel belt of the Alps-Betics orogen (in green).



The Apennines arc migrated eastward up to the present position in association with the slab retreat. The western Mediterranean back-arc basins developed in the hangingwall of this retreating slab. Note the lengthening of the arc which should generate also along arc extension. The western border of the back-arc rifting has a convex shape, mirroring the opposite Apennines arc. The initial rift cross-cuts the Alpine-Betic front, suggesting that the extension is not simply the collapse of the Alps-Betics orogen (after Doglioni *et al.*, 1999).

The rifting-drifting of Sardinia, and the formation of the Ligurian-Provençal Basin, as well as the emplacement of large amounts of igneous rocks have been related to the existence of the NW-directed Apennines subduction system that evolved to form the present-day Apennines belt (Carminati and Doglioni, 2005, and references therein). The geochemical composition of the Late Eocene-Middle Miocene igneous rocks of Sardinia resemble closely the magma presently generated along subduction settings, with abundant calcalkaline and high-K calcalkaline, with minor arc-tholeiitic compositions (Brotzu, 1997; Morra *et al.*, 1997; Mattioli *et al.*, 2000; Franciosi *et al.*, 2003; Lustrino *et al.*, 2004, 2009; Fig. 10b). Subduction-related coeval igneous rocks with a similar chemical composition were generated also along the Provençal coast (SE France), in SE-Corsica and offshore SW-Corsica. The volume and the areal extent of these products are about two orders of magnitude smaller than the volume of the Sardinian rocks. These features are probably related to the minor stretching rate in Provence and Corsica, located closer to the rotation hinge near the Gulf of Genoa (Fig. 1). During the 35-30 Ma interval, a nearly continuous NE-SW-trending volcanic line developed in the Western Mediterranean, with emplacement of subduction-related volcanic rocks in the Malaga area (S Spain; Duggen *et al.*, 2004), Mallorca island (Balearic Islands; Martì *et al.*, 1992), Sardinia (Lustrino *et al.*, 2009) and Provence (SE France; Beccaluva *et al.*, 2004). These rocks can be considered as the trace of the Apennines paleo-trench system (Lustrino *et al.*, 2011).

Figure 22. Within the Cenozoic evolution of the Mediterranean, Italy was shaped by the Alpine and Apeninnes subduction zones.



The early "east"-directed Alpine subduction was followed by the Apennines "west"-directed subduction,which developed along the retrobelt of the preexisting Alps. The slab is steeper underneath the Apennines, possibly due to the "westward" drift of the lithosphere relative to the mantle (after Carminati and Doglioni, 2005).

During Late Eocene-Middle Miocene times, the arc magmatism was almost exclusively confined to Sardinia, in the upper plate of the Apennines subduction. Other minor areas interested by subduction-related igneous activity are the Early Oligocene Po Plain in the Apennine foreland (Mortara buried volcano; ~30 Ma; Mattioli *et al.*, 2002a; Fig. 11c) and the nearly coeval (~29 Ma) conglomerates rich in volcanic fraction of northern Apennines (Aveto-Petrignacola Formation; Mattioli *et al.*, 2002b; Fig. 11c). The subduction-related igneous activity continued until Middle Miocene (~15 Ma; Fig. 11b-c) only in Sardinia. During Middle to Late Miocene (~15-7 Ma), no important igneous activity developed in the Italian area, whereas the westernmost Mediterranean was characterized by abundant subduction-related igneous activity in the Betics and Rif Belts (Turner *et al.*, 1999; Duggen *et al.*, 2004, 2005; Doblas *et al.*, 2007).

As outlined previously, starting from the latest Miocene - but essentially during Plio-Quaternary times - another phase of "subduction-related" igneous activity developed along peninsular Italy, from Tuscany (northern Apennines) to Campania (southern Apennines) regions, along the Tyrrhenian Sea border. This activity, almost entirely characterized by products with potassic to ultrapotassic composition, possibly witnesses the continental nature of the slab sinking beneath the Apennines. The origin of the potassic and ultrapotassic rocks of the Roman district (the so-called Roman Comagmatic Province of Washington, 1906) has been classically ascribed to partial melting of an heterogeneous mantle sources with phlogopite-rich veins originated after interaction of slab-derived melts and fluids with ambient mantle (Peccerillo, 1985, 1999, 2005; Beccaluva *et al.*, 1991; Conticelli *et al.*, 2002, 2004, 2009; Avanzinelli *et al.*, 2009, and references therein). In particular, the leucite-bearing magmas have been modelled with a process of interaction of sediments (essentially marls) brought at mantle depths during subduction processes (e.g., Peccerillo, 1985; Beccaluva *et al.*, 1991; Avanzinelli *et al.*, 2009, and references therein). In this case the mantle source should be lherzolitic to wehrlitic, in order to explain the CaO-rich composition of kamafugites and leucite-bearing ultrapotassic rocks. Experimental petrology studies evidence that melting of this type of lithologies under high X_{CO2} can produce liquids resembling kamafugites and other SiO_2-undersaturated ultrapotassic rocks even at relatively low pressures (see discussion in Avanzinelli *et al.*, 2009). The enrichment in Ca and CO_2, as well as the low SiO_2 content of the magmas, and the abundant CO_2 emission along the Italian peninsula and near the Calabria-Peloritani arc are considered to reflect interaction with $CaCO_3$-rich recycled sediments in a mantle wedge (e.g., Avanzinelli *et al.*, 2009; Frezzotti *et al.*, 2009, and references therein).

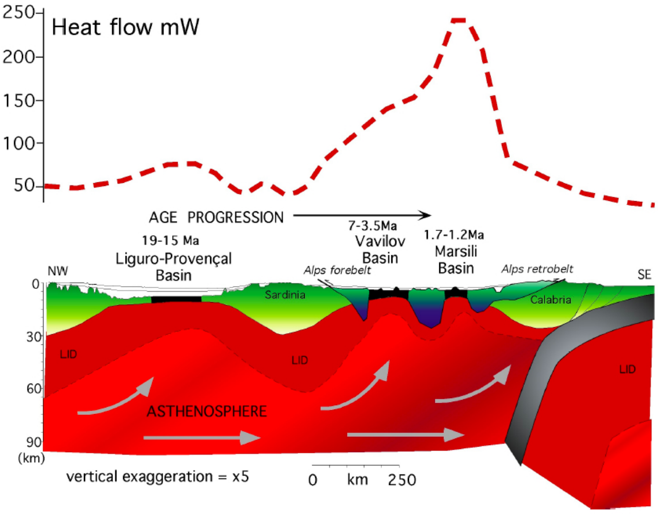
During the eastward retreat of the Apennines slab the upper plate, including the inherited Alpine orogen, has been sheared and boudinaged (Fig. 23). The retreat of the slab implies that asthenospheric mantle had to compensate the volume left by the migrating subduction. In other words, an eastward mantle flow has to be envisaged in the Western Mediterranean basin since the onset of the Apennines subduction, irrespective of whether it the cause of the slab retreat or the consequence. The shear-wave splitting analysis of this area (Lucente *et al.*, 2006) confirms a relevant seismic anisotropy that can be associated with mantle flowing from west to east in the Mediterranean and encroaching a slab, which constitute an obstacle deviating the flux (Fig. 24).

The boudinage and rifting in the Tyrrhenian Sea in the hangingwall of the subduction generated subsidence in the backarc, uplift above the subduction zone, and subsidence in the foredeep (Fig. 15). A number of different mechanisms controlled contemporaneously the vertical movements in the area. Lateral migration of depleted mantle can determine uplift; stretching and thermal cooling in the basin rather generates subsidence; mantle wedging and accretion above the subduction hinge produce uplift; slab rollback generates subsidence in the foredeep; loading of the prism and the foredeep trigger subsidence, etc. (Fig. 16). Therefore different geodynamic mechanisms may coexist along a cross-section showing that not a unique mechanism is determining the vertical movements, and that the subduction system may rather be considered as a passive process.

In this scenario, the down-flexure induced by the Apennines subduction (Fig. 15) explains the fast subsidence rates in the Po Plain and in the northern and central Adriatic Basin (Carminati *et al.*, 2003a).

The Pleistocene-Present uplift of the southern Adriatic Basin and of Apulia is explained by the entrance in the subduction zone of 110 km-thick continental lithosphere that chocked the subduction and induced buckling of the subducting lithosphere (Doglioni *et al.*, 1994). The fact that the highest mountains do not coincide with the water divide is interpreted as induced by the faster 'eastward' propagating tectonic wave generated by the retreat of the Apennines subduction zone during Pliocene and Quaternary times (10-30 mm/a) with respect to average denudation rates (<1 mm/a) (Figs. 12 and 13).

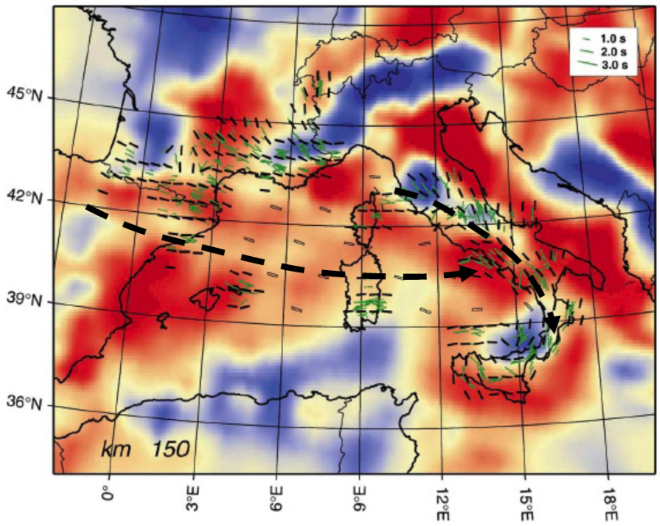
Figure 23. The western Mediterranean shows large asymmetric scale boudinage of the lithosphere in the back-arc extensional setting which developed particularly in the last 30-40 Ma.



The age progression from west to east is indicate by the eastward younging of the basins. The most active spredingis occurs the Tyrrhenian Sea where there is also the highest heat flow. The backarc extension in the Tyrrhenian Sea stretched the pre-existing Alpine double-verging orogen (dark blue and green). The "eastward" retreat of the Apennines slab has been compensated by the eastward mantle flow in the hangingwall. The oceanic crust is shown in black (after Gueguen *et al.*, 1997; Zito *et al.*, 2003).

Models involving nearly horizontal ruptures of a slab (slab break-off; e.g., De Astis *et al.*, 2006; Bianchini *et al.*, 2008), vertical ruptures of a slab (slab window; e.g., Schiattarella *et al.*, 2005; D'Orazio *et al.*, 2007) or more complex scenarios (horizontal detachment of subducted slab in post-collisional conditions coupled with vertically-propagating tears in the slab and differential slabs rollback; Rosenbaum *et al.*, 2008) have been alternatively proposed to explain the climax of the igneous activity in the Mt. Vulture area and in the Roman Province. In all cases, a toroidal mantle flow would have put in contact uncontaminated African (Adriatic) foreland mantle sources with subduction-related fluids and melts related to the Apennine subduction system, generating an hybrid mantle source with subduction-related geochemical features but also with still partially visible pre-metasomatic compositions, ranging from OIB- (Ocean Island Basalts) to DMM- (Depleted MORB Mantle) like; e.g., Peccerillo, 2005; D'Orazio *et al.*, 2007; Bianchini *et al.*, 2008; Rosenbaum *et al.*, 2008; Avanzinelli *et al.*, 2009).

Figure 24. Seismic anisotropy in the central-western Mediterranean and P-wave mantle tomography at the depth of 150 km (after Lucente *et al.*, 2006).



Note the alignment of the anisotropy (SKS splitting measurements) with the supposed mantle flow shown with the thick dashed black line in the hangingwall of the W-directed, easterly retreating Apennines slab. The inferred flow tends to parallel the slab when approaching the Apennines subduction zone, suggesting the presence of an obstacle able to deviate the mantle flow.

The tectonic setting of the Aeolian Islands is particular, being spatially and temporally associated to volumetrically important igneous activity in the neighbouring Mt. Etna, with emplacement of lavas with anorogenic geochemical characteristics (see discussion and references in Lustrino and Wilson, 2007). The peculiarity of this sector is based on the fact that the external area of the Aeolian Islands igneous activity (i.e., toward the trench) is occupied by the exotic terrane of Calabria (the tip of the Italian peninsula) and Peloritani Mts. (the NE-most margin of Sicily) which are separated from the Apennine-Maghrebian belt by the Sangineto and Taormina tectonic lineaments, respectively (e.g., Elter *et al.*, 2004, and references therein). The origin of the igneous activity in the Aeolian district is considered to be related to the metamorphic reactions of the Ionian subducting plate acting at depths ranging from ~100 to 250 km in depth. These slab dehydration processes metasomatize and soften the mantle wedge, lowering its solidus temperature, favouring partial melting processes (e.g., Peccerillo, 2005, Francalanci *et al.*, 2007; Chiarabba *et al.*, 2008; Peccerillo *et al.*, 2008, and references therein).

The origin of the volcanic activity of Mt. Etna shows many similarities with the Aeolian Islands, although the Aeolian volcanism is above the projection of the slab at depth (100-120 km), whereas the Etna volcano is located on the subduction hinge (Doglioni *et al.*, 2001). This feature is at the same time obvious and unexpected. It is obvious because Mt. Etna is located only few km south of the southernmost Aeolian island and because Mt. Etna and Aeolian Islands are nearly coeval. It is surprising because the products of the two volcanic areas have completely different geochemical and isotopic characteristics, incompatible with a similar mantle source. In particular, the formation of the Mt. Etna volcano has been considered to be related to an asthenospheric flow through a slab window created as consequence of differential roll-back of the subducted Ionian (Mesogean) oceanic lithosphere relative to the neighbouring Sicilian continental lithosphere. The Malta escarpment represents a Mesozoic-Tertiary passive continental margin separating the Sicilian and Ionian domains. This area acted as a transfer zone during the slab retreat of the Apennines, allowing larger roll-back of the Ionian oceanic lithosphere with respect to the Sicilian continental slab. The differential roll-back is evident when comparing the shallow dip of foreland regional monocline in the Hyblean Plateau with the steeper dip beneath the Ionian accretionary prism (Doglioni *et al.*, 2001). The tear separating the two different rates of rollback is concentrated along the inherited Malta escarpment, which was used during the Apennines subduction as a right-lateral transtension trasfer. This process allowed passive upwelling of asthenospheric mantle through such a vertical slab-window that generated partial melting as consequence of adiabatic decompression (e.g., Gvirtzman and Nur, 1999; Doglioni *et al.*, 2001; Armienti *et al.*, 2004). Such asthenospheric “suction” might also have mobilised subduction-related fluids coming from the neighbouring Ionian oceanic plate (e.g., Tonarini *et al.*, 2001).

The Hyblean Mts. igneous rocks show many geochemical, mineralogical and isotopic similarities with the Mt. Etna products (Figs. 10, 19 and 20), but their origin must rely on different geological processes, being the tectonic settings very different. Indeed, the recent igneous activity of Hyblean Mts. has been related to extensional tectonics and the development of pull-apart basins, with subsequent adiabatic decompression melting (Trua *et al.*, 1997; Beccaluva *et al.*, 1998). On the basis of major and trace element and Sr-Nd-Pb isotopic constraints, Beccaluva *et al.* (1998) proposed that the Miocene-Pliocene volcanism of the Hyblean Plateau was caused by partial melting of a depleted mantle source, variably modified by asthenosphere-derived melts or fluids, with both alkali-silicate and carbonatitic compositions, capable of stabilizing metasomatic phases (e.g., amphibole, phlogopite, apatite and carbonate), which partly replaced the original mantle paragenesis.

As the Tyrrhenian Sea igneous rocks are concerned, it must be noted that their major and trace element composition is certainly anomalous with respect to classical oceanic floor produced in divergent settings such as North Atlantic Ocean. This testifies to the occasional presence of metasomatic modifications of their sources related to slab-derived percolating melts/fluids. The average depth of the Tyrrhenian sea-floor is comparable with sea-floor depths of Atlantic Ocean of the same age (Kastens *et al.*, 1988; Korenaga and Korenaga, 2008).

Major and trace element content of Sicily Channel volcanic rocks is compatible with emplacement in a foreland rifted area from mantle sources that excaped subduction-related metasomatic modifications. More in detail, variable trace element ratios and elevated $^{207}Pb/^{204}Pb$ in Pantelleria rocks are believed to be indicators of the involvement of ancient enriched mantle source components (which resemble the enriched lithosphere sampled by circum-Tyrrhenian potassic magmas) in addition to DMM-HIMU (HIMU = High μ; $\mu = {}^{238}U/{}^{204}Pb$) (± EMI = Enriched Mantle type I) mantle sources (Esperança and Crisci, 1995; Civetta *et al.*, 1998). With the exception of one sample, the total range of $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ is 0.7030-0.7033 and 0.51300-0.51292, respectively (Fig. 17b). Compared to the limited range of Sr-Nd isotope compositions, the Pb isotopic ratios show a much wider range ($^{206}Pb/^{204}Pb$ = 18.30-19.94, $^{208}Pb/^{204}Pb$ = 38.04-39.62: Δ7/4 = -6.4 to +11.6; Δ 8/4 = -39.4 to +45.8; Fig. 19b). Trace element abundances of Linosa Island volcanic rocks (Rossi *et al.*, 1996; Bindi *et al.*, 2002; Peccerillo, 2005) show the typical bell-shaped patterns in primitive mantle-normalized multi-elemental diagrams, peaking at Nb with relatively low LILE abundances, within the HIMU-OIB compositional field (Fig. 20b). Sr-Nd-Pb isotopic ratios range as follows: $^{87}Sr/^{86}Sr$ from 0.7030 to 0.7031, $^{143}Nd/^{144}Nd$ from 0.51298 to 0.51295, $^{206}Pb/^{204}Pb$ from 19.28 to 19.43, $^{207}Pb/^{204}Pb$ from 15.60 to 15.62, $^{208}Pb/^{204}Pb$ from 38.89 to 39.02, Δ7/4 from +1.5 to +2.8, Δ8/4 from -12.5 to -4.3 (Figs. 17b and 19b). Esperança and Crisci (1995) proposed a petrogenetic model in which a lithospheric mantle source, enriched and metasomatized by asthenospheric partial melts, undergoes partial melting in an extensional tectonic regime without any mantle plume involvement. Corti *et al.* (2006) propose an origin for Pantelleria and Linosa islands related to the oblique crustal extension in the foreland of the Apennines-Magrebide front, testifying the existence of two completely different and unrelated geodynamic processes (continental extension and active subduction). This coexistence is very relevant in supporting the idea of the passive nature of plate boundaries (i.e., both their geometry and evolution are governed by forces associated to far processes), casting serious doubts on their relevance in driving plate tectonics. Volcanic rocks from Graham and Nameless Banks in the Sicily Channel have geochemical characteristics similar to Pantelleria and Linosa Islands, broadly resembling HIMU-OIB. Small differences are related to sea-water alteration (Rotolo *et al.*, 2006). As a whole, the Sicily Channel seamounts have $^{87}Sr/^{86}Sr$ ranging from 0.70308 to 0.70478, $^{143}Nd/^{144}Nd$ from 0.51312 to 0.51295 (Fig. 17b), $^{206}Pb/^{204}Pb$ from 19.15 to 19.81, $^{207}Pb/^{204}Pb$ from 15.66 to 15.70, $^{208}Pb/^{204}Pb$ from 38.95 to 39.62, Δ7/4 from +3.9 to +13.0, Δ8/4 from -13.0 to +19.4 (Fig. 19b).

Ustica lavas have a small range in $^{87}Sr/^{86}Sr$ (0.7030-0.7033) and $^{143}Nd/^{144}Nd$ (0.51300-0.51290; Fig. 17b) and a slightly wider range of compositions for $^{206}Pb/^{204}Pb$ (18.84-19.56), $^{208}Pb/^{204}Pb$ (38.61-39.19), Δ7/4 (+3.3 to +7.4) and Δ8/4 (-7.8 to +20.8); Trua *et al.*, 2003; Peccerillo, 2005; Fig. 19b). On the basis of their trace element and Sr-Nd isotopic compositions, Cinque *et al.* (1988) and Trua *et al.* (2003) proposed that the source of the Ustica alkali basalts was slightly contaminated by a component of slab origin. It has been suggested that the volcanic activity of Ustica (together with that of the Prometeo seamount and Mt. Etna) can be explained without the need to invoke a deep-rooted mantle plume, but considering instead the flow of African mantle below Sicily, channelled along the SW edge of the Ionian oceanic plate which is undergoing subduction roll-back (e.g., Gvirtzman and Nur, 1999; Doglioni *et al.*, 2001; Trua *et al.*, 2003).

Alternative geodynamic models for the Apennines: a discussion

As discussed, the coupled Oligocene to Present extensional-compressional wave in the Apennines-Provençal-Tyrrhenian system and the associated magmatism is related by a large number of authors (e.g., Scandone, 1979; Dercourt *et al.*, 1986; Malinverno and Ryan, 1986; Beccaluva *et al.*, 1989; Doglioni *et al.*, 1997; Carminati *et al.*, 1998; Gueguen *et al.*, 1998; Faccenna *et al.*, 2001) to the radial retreat of a NW-directed subduction of oceanic lithosphere beneath the European southern paleo-margin. However the same geological and igneous petrological observations were ascribed alternatively to different hypotheses such as: a) mantle upwelling (called “asthenolith” or “mantle megadome” in the literature) impinging the south-eastern European lithosphere around Late Oligocene (e.g., Wezel, 1982; Burrus, 1984; Doblas and Oyarzun, 1990; Lavecchia and Creati, 2006; Centamore and Rossi, 2008), migrating in W-E direction from the Alboran to the Aegean basin from Oligocene to the Present (Morelli, 1998); b) lateral expulsion of crustal wedges (e.g., Carmignani *et al.*, 1994, Assorgia *et al.*, 1995; Pasci, 1997; Mantovani *et al.*, 2002); c) delamination of a lower crustal and lithospheric mantle slab, whose tear migrates through time to the NE (e.g., Channell and Mareschal, 1989). In some cases the geochemical characteristics of the Apennines Plio-Quaternary magmatism have been considered unrelated to any type of subduction processes (e.g., Scalera, 2006), but rather generated by an alternatively hot (e.g., Bell *et al.*, 2006) or “wet” (with “normal” temperature; Lavecchia and Creati, 2006) uprising mantle. In particular, Bell *et al.* (2006) propose the existence of a large mantle plume, with a diameter of at least 1000 km from Sardinia to the NW Alps and peninsular Italy. The existence of a large range of chemical and isotopic compositions among the Plio-Quaternary Italian igneous rocks is simply related to a chemically and isotopically heterogeneous alleged mantle plume. Lavecchia and Creati (2006) state that “the Apennines and the Maghrebides are not related to subduction, but represent a recent example of «plume-induced orogenesis»”. In their view, the seismically active Benioff-Wadati plane beneath the Calabro-Peloritani arc would represent the strongly deformed (originally SE-dipping) Alpine slab (Lavecchia and Creati, 2006). Also Locardi and Nicolich (2005) proposed the existence of an eastward migrating deep mantle plume (called asthenolith) to explain the opening of the Ligurian-Provençal and Tyrrenian basins, their reduced lithospheric thickness, the rotation and collision of several microplates in the Western Mediterranean and the igneous activity. All these papers are essentially based on the original model proposed by Wezel (1982). In other cases a mix model, requiring the existence of both subduction processes and a mantle plume has been proposed in literature. Gasperini *et al.* (2002) and Cadoux *et al.* (2007) explained the geochemistry of the peninsular Italy rocks with the involvement of a mantle plume uprising from deep mantle just in a hole of the subducting Apennines slab.

Also the igneous rocks of Sicily or Sicily Channel have been related to the existence of active or fossil deep mantle plumes. On the basis of detailed global seismic tomographic considerations, Montelli *et al.* (2004) suggested the presence of a deep mantle plume beneath Mt. Etna, similarly to what proposed by Hoernle *et al.* (1995) at a much larger scale, including the entire Mediterranean Sea and neighbouring continental areas. On the basis of major and trace element and isotopic constraints, Civetta *et al.* (1998) proposed that the magmatism of Pantelleria, as well as that of Mt. Etna, is related to the activity of a mantle plume. Rotolo *et al.* (2006) instead proposed the presence of a fossil plume head beneath the Sicily Channel, Mt. Etna, the Hyblean Plateau, Ustica Island and the westernmost island of the Aeolian Archipelago (Alicudi). Their model is based on the relatively homogeneous EAR-like (EAR = European Astenospheric Reservoir; Granet *et al.*, 1995) major and trace element and Sr-Nd-Pb isotopic characteristics of the magmatic rocks.

We believe that there is no necessity to relate the igneous activity in these areas to the existence of active or fossil upraise of solid mantle from very deep. These are often unsupported by geophysical, thermal, sedimentological and geological data. Also geochemistry (commonly considered the strongest evidence for the existence of such type of processes) can be explained in different ways, without invoking such large-scale events (Lustrino and Wilson, 2007, and references therein). More detailed considerations on alternative hypotheses proposed to reconcile the chemical-isotopic composition of Plio-Quaternary igneous rocks of Italy with geodynamics are fully approached in Lustrino *et al.* (2011).

Differences between Alps and Apennines

As outlined above and discussed by Carminati *et al.* (2004a), Alps and Apennines show distinct characters (e.g., Figs. 7, 8, 9, 17). They have, respectively: 1) subduction hinge moving towards vs. moving away from the upper plate; 2) double vs. single vergence; 3) high vs. low morphological and structural elevation; 4) deep vs. shallow basement rocks involved; 5) higher metamorphic grade vs. lower metamorphic grade; 6) the basal décollement involves the crust and the LID of both upper and lower plates in the Alps whereas only the shallow crust of the lower plate contributes to the accretionary prism in the Apennines; 7) shallow vs. deep foredeep; 8) low vs. high dip of the foreland monocline; 9) thickened vs. thinned crust under the ridge; 10) the Alps have both in the upper and in the lower plate a pre-subduction Moho, whereas the Apennines have in the footwall plate a pre-subduction Moho, but in the upper plate they present a new forming Moho; 11) thickened lithosphere vs. a shallow asthenosphere in the upper plate; 12) no back-arc basin and syn-subduction magmatism vs. very wide back-arc basins and abundant syn-subduction arc-tholeiitic to calcalkaline and potassic/ultrapotassic magmatism; 13) >99% of the igneous products in plutonic facies, with relatively important shallow crustal interaction of mantle melts in the Alps vs. >99% of the igneous products in effusive to pyroclastic facies and essentially mantle-derived melts in the Apennines and Apennine-related igneous activity; 14) SiO_2-oversaturated vs. SiO_2-saturated to strongly SiO_2-undersaturated compositions of igneous rocks; 15) low vs. high gravity and heath flow anomalies (in particular in the internal part of the Apennines). These changes reflect, at least in part, the first order differences between “west” and “east” directed subduction zones, as widely discussed by Doglioni *et al.* (1999; 2007).

The Italian orogens represent two prototypes of subduction zones whose main characters can be exported throughout the Mediterranean and the entire Earth. The basic difference is the asymmetric behaviour of the decoupling zones. The fundamental decoupling is the one separating lithosphere and the underlying mantle, occurring in the low-velocity layer atop of the asthenosphere. Along W-directed subduction zones, this decoupling is folded and subducted, being the slab dragged eastward by the relative motion of the mantle, since the lithosphere has a net westerly directed rotation (Crespi *et al.*, 2007). Along E- or NE-directed subduction zones, i.e., along the slabs following the mantle flow in the direction of the tectonic equator, the basal decoupling is rather uplifted toward the surface, allowing exhumation of deep seated rocks, and wide shortening of the upper plate. In other words, the main difference between the two subduction end-members is the trajectory of the decoupling between lithosphere and mantle. Moreover, where the subduction hinge is retreating relative to the upper plate as it is in the Apennines, there is not an upper plate actively converging over the lower plate, which is instead the case of the Alps.

Conclusions

The geodynamics of the Italian area are part of the larger Mediterranean evolution. The movie included in this article tries to synthetize our interpretation on the geodynamic and magmatic evolution of the Mediterranean during the Tertiary. The Alpine subduction generated a double-verging elevated orogen and developed from the Early Cretaceous to Present. The Apennines nucleated along the retro-belt of the Alps, from the Marittime Alps in western Liguria moving southward, where oceanic or thinned continental lithosphere was present (Fig. 23). The Apennines subduction had the subduction hinge migrating away relative to the European upper plate, and hence generated the back-arc basin that constitutes the entire western Mediterranean (Fig. 24). The back-arc rift was associated to asymmetric boudinage. The Apennines subduction retreat generated an arc characterized by second order undulations, i.e., salients and recesses that were controlled by the inherited lithospheric structures. The largest salient occurred in the Calabria arc and its offshore accretionary prism propagation, where there was the most subductable foreland, i.e., the likely oceanic Ionian/Mesogean Basin. The most prominent recess rather occurred at the intersection of the migrating Apenninic arc with the thick continental Adriatic lithosphere, which is resisting subduction since Pleoistocene, possibly due to its more effective buoyancy. The southern branch of the arc in Sicily is presently subject to the subduction hinge moving toward the upper plate (Devoti *et al.*, 2008). This generated an inversion of the system; as a result, the northern Sicily offshore-southern Tyrrhenian Sea is now under compression, in spite of being located in the extending backarc basin (Scrocca *et al.* 2008). This inversion likely occurred since Pleistocene.

The Alpine and Apennines Belts were accompanied by igneous activity since Paleocene. The most ancient products are volcanic and sub-volcanic rocks emplaced in the Veneto Region (NE Italy; ~65-25 Ma), Punta delle Pietre Nere (Gargano peninsula; central-eastern Italy; ~62-56 Ma), Mt. Queglia (Abruzzi region, central Italy; <54 Ma) and Sulcis district (south-west Sardinia; ~61-51 Ma). These products have major and trace element content as well as Sr-Nd-Pb isotopic ratios compatible with a derivation from mantle sources not sensibly modified by subduction-related processes (i.e., are considered to be “anorogenic” or “intra-plate-like”). Other igneous products with roughly similar compositions are emplaced much later mostly in north-east Sicily (Mt. Etna; <1 Ma), in the Sicily Channel (Pantelleria and Linosa Islands plus minor seamounts; mostly <1 Ma), in south-west Tyrrhenian Sea (Ustica Island; <1 Ma). In south-eastern Sicily a long-lasting igneous activity with “anorogenic” geochemical characteristics is recorded in the Iblean Mts. area. The last area where “anorogenic” volcanic rocks crop out is Sardinia. Here detailed isotopic investigation allowed to distinguish two different igneous phases, the first dated late Middle Miocene-Early Pliocene (~12-4.4 Ma), the second dated Early Pliocene-Quaternary (~3.9-0.1 Ma). The products of the first phase crop out only on the southernmost sectors of the Island and constitutes <<1% of the volcanic rocks, wherease the products of the second phase crop out in central-northern Sardinia and are by far the most exotic composition ever recorded in the entire circum-Mediterranean area. To summarize, “anorogenic” magmatism has been active in the Italian area from ~65 to ~25 Ma and from ~12 Ma to Present. With the exception of the igneous activity in north-east Italy (Veneto Region) no “anorogenic” igneous rocks have been produced during the 51 to 12 Ma interval.

The volume of “subduction-related” or “orogenic” igneous rocks are three to five orders of magnitude larger than the volumes of “anorogenic” magmas produced during the Cenozoic in the Italian area. Most of the “orogenic” igneous activity is concentrated in three periods: Early Oligocene (~33-27 Ma) along the Insubric Line in the Alps, Early Miocene (~22-18 Ma) in Sardinia (and possibly in the Ligurian-Provençal Basin), and Plio-Quaternary (<2 Ma) along peninsular Italy and in south-eastern Tyrrhenian Sea (Aeolian Islands). The major and trace element compositions, as well as key isotopic systematics, are different in the three areas, depending on the thermal state, the composition and the age of the local mantle(s), the composition and the age of the subducting material and the styles of interaction between subduction-related metasomatic agents and the peridotitic matrix. In some cases a sort of geochemical inheritance of the mantle sources seems to have had a stronger effect in determining the final composition of the magmas compared to the potential modifications related to geological processes coeval to the igneous activity.

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